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GEOLOGICAL MAGAZINE

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Some Aspects of the Alpine-Type Serpentinites of Queensland

By J. F. G. WILKINSON

(PLATE XII)

ABSTRACT

Several aspects of the Queensland serpentinites are discussed, including the significance of a suite of serpentinite-breccias, the formation of antigorite, and the general problem of serpentinization and mode of emplacement of the ultramafic rocks.

INTRODUCTION

ALTHOUGH ultramafic rocks have been recognized for many years in Queensland, previous references to them have been concerned mainly with brief reports on economic mineral associates, and scant attention has been directed to their petrology. The main purpose of this paper is to summarize the principal features of the Queensland serpentinites,[1] arising from a study of these rocks at intervals in the years 1949–1950, in which several months were spent in the field and nearly 400 thin-sections of the various rock types prepared. More detailed accounts of the ultramafic and associated rocks are to be embodied in publications in Australian journals.

Since 1951 the writer has had the opportunity of examining various areas along the Great Serpentine Belt of New South Wales, made classic by the researches of Benson (1913–18). This later work has tended to strengthen certain conclusions previously drawn from a study of the Queensland occurrences. Comparison of the ultramafic and associated rocks (both igneous and sedimentary) from both States, has shown a striking similarity in field relationships, tectonic and structural background, petrology, chemical characteristics, and rock associations in widely remote regions. Such features are best explained by regarding the ultramafic rocks of Queensland and New South Wales as part of the same belt, which could be referred to as the Great Serpentine Belt of Eastern Australia. In the following discussion

[1] A term used for rocks composed of minerals of the serpentine group. Subsequent laboratory investigation may designate a serpentinite as an antigorite as the case may be (cf. Selfridge, 1936, p. 501).

on the Queensland rocks, many of the features are therefore duplicated in the New South Wales occurrences. It is hoped to commence a detailed study of certain aspects of the mineralogy of the ultramafic rocks at a later date.

Distribution of the Ultramafic Rocks, Form of Intrusions, and Geological Setting

The known occurrences of ultramafic rocks in Queensland are shown on the accompanying sketch-map (Text-fig. 1). The intrusions at Pine Mountain, Fernvale, Northbrook, and in the Mary Valley and Cawarral-Tungamull areas, were examined in detail, whilst work of a reconnaissance nature was carried out elsewhere. Present knowledge indicates that the length of the belt in Queensland, from Pine Mountain in the south of the State, to Temple Bay, to be of the order of 1,200 miles.

The ultramafic intrusions vary considerably in size, from elongate lensoidal outcrops up to several hundred yards in length, to larger bodies, 20 to 30 miles long with a width of 4 to 5 miles, and usually reveal a general concordancy with the intruded rocks, although locally apophyses may cut adjacent Lower Palaeozoic sediments. The dominant strike of the intrusions is generally north-north-west to north-west, so that they approximately parallel the axis of the major Palaeozoic structure of Eastern Australia, namely the Tasman Geosyncline, whose history was one of varied sedimentation and igneous activity from the Pre-Cambrian to the end of the Permian. An even more striking parallelism of the ultramafic rocks is afforded by their localization along zones of strong dislocation, e.g. the presence of serpentinites along the plane of the West Moreton Fault system at Pine Mountain, Fernvale, and Northbrook.

At Pine Mountain, amphibolites occur along the thrust in association with the serpentinites, in the midst of a terrain of unmetamorphosed rocks (cherts, jaspers, greywackes, etc.). These amphibolites are regarded as thrust-slices from a high grade Pre-Cambrian basement, and are thought to have been brought to their present position as a result of those strong compressive movements during which the contents of the geosyncline were folded, and intrusion of the ultramafic rocks occurred. These metamorphic rocks are not modifications of the serpentinites, or products of the higher temperature type of metamorphic differentiation at serpentinite-siliceous country-rock contacts (Phillips and Hess, 1936, p. 341).

The ultramafic rocks under discussion belong to that group found along the major axes of island arcs and alpine mountains (the " alpine type " of Benson, 1926), and possess those features characteristic of the primary peridotite suite of Hess (1937*b*, 1938). They are therefore

distinct from those ultramafics of the basaltic magma suite, e.g. ultramafic types present in the lower levels of certain teschenitic sills and gabbroidal-noritic lopoliths.

TEXT-FIG. 1.—Sketch-map showing the distribution of the serpentinites in Eastern Australia. (1) Pine Mountain, Fernvale, and Northbrook. (2) Mary Valley. (3) Mundubbera, Yarrol, Many Peaks. (4) Cawarral-Tungamull. (5) Princhester-Marlborough. (6) Mount Pring near Bowen. (7) Gray Creek. (8) Temple Bay. (9) Near Eddystone Vale. (10) Theresa Creek. (11) The Great Serpentine Belt of New South Wales.

(Note that some of the smaller intrusions have been enlarged. Also that certain occurrences may not be entirely continuous.)

In Queensland the most frequent field associates of the serpentinites are rocks belonging to the Neranleigh-Fernvale Group of the Brisbane Metamorphics, into which the ultramafics have been intruded. Though typically developed in the vicinity of Brisbane, representatives of the Group are recognized at considerable distances from the type area, e.g. in the Cawarral-Tungamull area, near Rockhampton. A generalized table of the Brisbane Metamorphics is given below (mainly after Bryan and Jones, 1950).

TABLE 1

European Record	Stratigraphical Term		Representative Rock Types
Silurian	Neranleigh-Fernvale Group.	Brisbane Metamorphics	Greywackes, subgreywackes, cherts, jaspers, phyllites, andesites, limestones, calc-epidote rocks.
Ordovician	Bunya Phyllites		Mica phyllites, quartz-mica schists, plus psammitic types.
Cambrian	Rocksberg Greenstones		Low-grade metamorphosed andesites and basalts.

As yet the Neranleigh-Fernvale rocks have proved barren of any diagnostic fauna; they possess those characteristics generally ascribed to geosynclinal sedimentation. Spilites as field associates of the serpentinites, cherts, and jaspers, have not been recognized in Queensland. This relationship is known in New South Wales, e.g. in the Nundle area, near Tamworth. Although jaspers (often radiolarian) frequently occur in the field with serpentinites, it is not thought that any genetic relationship exists between the two rocks, since jaspers may be present remote from cherts which are themselves marginal to the serpentinite contacts, e.g. in the Mary Valley area.

In some instances rocks considerably younger than the Neranleigh-Fernvale Group are intimately associated with the ultramafics. Thus at Pine Mountain, Middle Triassic conglomerates, sandstones, shales, and tuffs of the Ipswich Coal Measures occur along the western edge of the intrusion. These rocks have been brought into their present position by post-serpentinite movements along the major fracture. Such posthumous movements have done much to complicate accurate dating of the ultramafic activity.

PETROGRAPHY OF THE SERPENTINITES

Both field and laboratory investigation have shown that the serpentinites of the Palaeozoic orogenic belt of Eastern Australia may

be considered in terms of three groups, each group possessing distinctive megascopic and microscopic characteristics.

I. *Massive Serpentinites.*—These rocks, which can be referred to various precursors of the peridotite suite, invariably reveal complete serpentinization, and under the microscope are characterized by the mesh structure of chrysotile serpentine after olivine. Evidence of the existence of earlier olivine is available even if transformation to antigorite occurs. Cross-fibre chrysotile is the common asbestiform mineral.

In the Queensland massive serpentinites, the presence of relict structures and minerals indicates that these rocks were originally harzburgites and dunites. Lherzolitic types have also been recognized from the Mary Valley. No relict olivine has been detected in any of the Queensland rocks. Where present, orthopyroxene is usually converted to pseudomorphs of bastite, though occasional enstatite cores do remain. The clinopyroxene remains quite fresh. Magnetite, chromite, and picotite are common accessory minerals. A feature of some slices is the presence of myrmekite-like intergrowths of picotite and orthopyroxene, now mainly bastite. The veins of cross-fibre chrysotile postdate all other mineral parageneses. No evidence of pyrogenetic amphibole, mica or feldspar has been observed in any slice of these rocks, a statement that is equally applicable to the rocks of Groups II and III.

II. *Schistose Serpentinites.*—These rocks possess both megascopic and microscopic schistosity and are notable under the microscope for the complete absence or paucity of structures which would result from the serpentinization of pre-existing minerals. They consist almost entirely of structureless chrysotilic serpentine, along with accessory magnetite, chromite, and picotite. Slip-fibre chrysotile is the common asbestiform mineral. In certain intrusions, the schistose serpentinites are equal to or may predominate in areal outcrop over the massive varieties.

III. Critical relationships between these distinct types of serpentinite exist in the occurrence of *composite rocks*, characterized by inclusions of massive serpentinite in a schistose matrix. The converse relationship has not been observed. These inclusions range progressively from rounded blocks of serpentinized harzburgite and dunite several feet in diameter, through inclusions the size of one's fist, to rocks consisting of angular chips and fragments in a serpentinous matrix, namely serpentinite-breccias (Plate XII, fig. 1). Angularity of the inclusions increases with decreasing size; the larger inclusions over 6 inches in diameter are generally well-rounded and are quite distinct megascopically and microscopically from the enclosing schistose serpentinite. Schistosity is developed parallel to the edges of the inclusions and wraps around curved faces. The junction between both rock types

is quite clean and distinct, without any indication of annealing or incorporation. In the finer-grained breccias the angular fragments may range in size down to 0·05 mm. in diameter (Plate XII, fig. 2), so that under crossed nicols a minute fragment is seen to be represented by portions of a single mesh.

Rocks illustrating the above relationships occur in intrusions from near Rockhampton, to the Port Macquarie and Lower Manning serpentinites in New South Wales. A linear distribution of the breccias has been observed in the Mary Valley. In some parts of an intrusion, the schistose phase may carry few inclusions; elsewhere, abundant closely packed fragments may exceed the serpentinous matrix in volume.

TABLE 2

	1	2	3	4 *	5
SiO_2	40·00	47·40	40·10	38·23	39·66
TiO_2	0·03	0·02	0·02	0·04	—
Al_2O_3	1·09	0·43	1·83	0·43	1·15
Fe_2O_3	8·21	6·12	6·04	4·27	7·66
FeO	1·22	0·99	2·32	1·73	0·98
MnO	0·14	0·07	0·10	0·21	—
MgO	34·20	29·90	35·60	38·99	36·20
CaO	1·20	0·50	0·49	1·09	0·26
Na_2O	tr.	tr.	tr.	nt. fd.	0·11
K_2O	tr.	0·07	0·05	nt. fd.	0·01
H_2O+, H_2O-	13·60	14·22	13·23	13·35	13·45
P_2O_5	0·01	0·01	0·01	—	—
Cr_2O_3	0·41	0·41	0·38	0·36	0·35
NiO	0·33	0·39	0·46	0·06	—
Total	100·44	100·53	100·63	98·76	99·83

* 2·57 per cent CO_2 omitted.

Analysis 1: Serpentinized harzburgite inclusion, Mary Valley, Queensland. Analyst: T. R. Lowth, Govt. Chem. Lab., Brisbane.

Analysis 2: Schistose serpentinite at contact of inclusion in Analysis 1. Analyst: T. R. Lowth.

Analysis 3: Serpentinite-breccia, Mary Valley, Queensland. Analyst: T. R. Lowth.

Analysis 4: Antigorite-serpentine, Bowling Alley Point, from the Great Serpentine Belt of New South Wales. (Benson, 1913, Part 3, p. 705.)

Analysis 5: Average of four analyses of the ultramafic series of Cuba. (Hess, 1938, p. 330.)

Chemical analyses of selected specimens of the three varieties of serpentinite are given in Table 2. Microscopic examination of specimen 1 shows it to be typical of the massive serpentinized harzburgites. The schistose serpentinite reveals unusually high SiO_2 and low MgO percentages. Analysis 3 is from a typical serpentinite-breccia; the largest fragments are of the order of 1 centimetre. The complete serpentinization suffered by the ultramafic rocks is reflected in their high

combined water content. Except for the silica and magnesia in analysis 2, a striking similarity is afforded by comparison with Analyses 4 and 5.

The Significance of the Inclusions and Breccias

In the Eastern Australian serpentinite-breccias the general absence of material foreign to the peridotite suite, and their distribution, namely in linear intrusions rather than pipes or necks, serves to distinguish them from ultramafic breccias elsewhere, e.g. the kimberlite of South Africa. The chemical composition (Table 2, Analysis 3) is in accord with both field and microscopic evidence of fragmental serpentinized peridotite in a serpentinous matrix.

Bowen (1949, p. 455) has advocated that peridotites, when not differentiated in place, have been intruded essentially in the solid state. The presence of inclusions and breccias might therefore imply that they are a natural accompaniment to solid intrusion. Benson (1913, Part II, p. 583) suggested that the schistose phase has been developed in situ from the massive variety by post-consolidation shearing stresses. The evidence which follows is deemed to be inconsistent with these concepts.

(i) The inclusions had already undergone normal magmatic consolidation before incorporation in the schistose phase. As yet, the only instance of primary ultramafic banding has been found at Port Macquarie in New South Wales, where a large inclusion, 15 feet in diameter, consists of alternating layers of serpentinized peridotite and enstatolite. The enstatite crystals, which may attain a length of 4 inches, are now converted to bastite. In addition, inclusions may be veined with cross-fibre chrysotile or show partial steatitization, although these features are absent from the enclosing schistose serpentinite.

(ii) In the one outcrop of breccia the serpentinized inclusions may vary in nature from dunitic to lherzolitic. Fragments of antigorite, chrysotilic mesh-serpentine showing transformation to antigorite, and unaltered mesh-serpentine may all be present in a thin-section of breccia.

(iii) In some breccias the volume occupied by the fragments (which may exhibit considerable size variation in the one hand specimen) is considerably greater than that of the matrix.

(iv) In an intrusion the schistose phase is not localized at the margins, but may be present in any position in the intrusion with respect to the serpentinite contacts.

(v) With increasing size of fragments there is a progressive increase in the degree of rounding.

The above facts, allied with the clean junction between inclusions and matrix, are taken to indicate that the inclusions were solid at a

time when the matrix was fluid or extremely plastic. The foregoing discussion also precludes the possibility of the inclusions and breccias being the result of violent eruption, in the manner suggested by Hess (1938, p. 328).

In the writer's opinion the various relationships existing between inclusions, breccias, and matrix can be explained by the conception of two phases of ultramafic activity along the fracture system where the rocks are now found. This is not unexpected since this major line of structural weakness has played a significant rôle in the localization of much of the post-serpentinite igneous activity. The first injection gave rise to normal serpentinized peridotitic types; at certain levels consolidation may have taken place under relatively tranquil conditions as indicated by the presence of primary banding. At a later phase in the geosynclinal history, a second orogenic episode led to further movement along the original dislocation. The massive serpentinites in many instances were fractured and brecciated, and the second phase, now represented by the schistose serpentinites, moved up and around the massive types and caught up blocks of the latter, so that present-day relationships are now realized. The twisted and contorted nature of the second phase in many places may indicate that the orogeny attendant upon its intrusion was the more important in intensity. Immediately following fracture and brecciation, the included blocks were probably angular, but in their subsequent movement in the schistose phase to their present positions, they became rounded by mutual abrasion, perhaps in certain instances before actual injection of the second phase. This progressive rounding of the larger fragments produced a variable series of smaller angular fragments, in the last recognizable instance being represented by microscopic angular chips of mesh-serpentinite (Plate XII, fig. 2). It should be mentioned that the serpentinite-breccias in the Mary Valley are closely associated with certain Kandanga Metabasites (see p. 315) which have suffered brecciation as a result of renewal of movement along the fault zone. Such rocks occur in the centre of the serpentinite belt.

In the light of the foregoing discussion, it would appear that the Australian occurrences constitute a noteworthy exception to the latter half of the statement by Hess (1937*a*, p. 269; 1948, p. 438) that orogenic ultramafics "are intruded during the first great deformation . . . they do not accompany later deformations".

The Age Problem of the Serpentinites

For many years it has been recognized that certain ultramafic types are characteristic of certain tectonic environments. Modern views are contained in a recent statement by Hess (1948, p. 432), namely: ". . . peridotites of the ultramafic magma suite occur in all

alpine type mountain systems and nowhere else. They occur in two belts about 50 miles on either side of the original location of the tectogene axis . . . They are intruded during the first great deformation of the belt, presumably during buckling of the crust, and later deformations of the same belt are not accompanied by intrusion of peridotites. Thus location of the peridotite belt and dating its intrusion locate the old tectogene axis and date the initiation of the deformation of that zone." The accompanying sketch-map (Text-fig. 1) indicates that the distribution of the orogenic serpentinites in Eastern Australia conforms approximately to the pattern outlined by Hess. (In Queensland the writer has only studied occurrences of serpentinite from Rockhampton to Pine Mountain, but it is likely that the other occurrences belong to the ultramafic belt in question.[1])

A consideration of the age of intrusion of the ultramafics is thus essential to an understanding of the tectonics of the Tasman Geosyncline. That it represents a difficult problem is indicated by the various ages assigned to the rocks, e.g. Devonian (Bryan and Jones, 1945), Carboniferous (Benson, 1913, Part I), and Permian (Voisey, 1939). Present data suggests that future generalizations with respect to the serpentinites be made after critical analysis of evidence available at different points in the belt. Rocks representative of a considerable span of geological time are intimately associated with the ultramafics. It is not intended here to enter into a lengthy discussion of the evidence for the various ages; rather it is intended to point out some of the difficulties involved in age determination. In many areas the associated rocks possess complex tectonics and there is frequently a dearth of determinative fossil evidence. Furthermore, the ultramafics tend to differ from most conventional igneous activity in their absence of contact metamorphism towards the intruded rocks, and in their location on deep-seated thrust planes, along which subsequent movement has frequently brought younger rocks into contact with the serpentinites.

It is generally accepted that three powerful orogenic movements can be recognized in the Tasman Geosyncline, namely Middle Devonian, Lower Carboniferous, and Late Permian. The latter orogeny saw the end of the Tasman Geosyncline as such. From a study of the tectonics of the Geosyncline in Queensland, D. Hill (1951, p. 16) places its first great climacteric between the Lower or lower Middle Devonian and the upper Middle Devonian, and she associates the

[1] Future detailed work on some of the western ultramafic intrusions in Queensland, e.g. near Eddystone Vale and Theresa Creek, may show that they are older than the main belt to the east. Thus ultramafic rocks at Gundagai, Coolac, and Young which lie to the west of the Great Serpentine Belt of New South Wales, are regarded as Silurian in age.

intrusion of " most of the great serpentine belts of Queensland " with this climacteric. Such an association is in accord with the earlier statement of Hess.

But another possibility arises from the widespread occurrence of serpentinite-breccias in the belt, for such relationships imply a time-break between intrusions. From the Wood's Reef serpentinite, in the Great Serpentine Belt of New South Wales, Osborne (1950, p. 69) and Proud and Osborne (1952, p. 17) recognize " intrusion-breccias " and two injections of ultramafic material, the first dominantly harzburgitic, followed by a dunitic phase. Whilst also recognizing a definite time-break between the intrusions, the above writers regard this as relatively short and refer both phases of igneous activity to late Permian times. However, at the southern end of the Mary Valley serpentinite in Queensland, where massive serpentinites (now dominantly antigoritic) predominate over the schistose varieties, inclusions of antigoritic serpentinite and serpentinite-breccias containing antigorite fragments, are found in a matrix of normal chrysotilic serpentine. It is likely then that the time interval between intrusions may have been appreciable, since in this area the intrusions are separated by an intervening metamorphic episode. Future reconciliation on the problem of age may therefore be possible, e.g. intrusion along the same lines of weakness in both Middle Devonian and Lower Carboniferous times. Such localization of injection is not unlikely when the control of the major fracture system on much of the post-serpentinite igneous activity is considered.

The following lines of research may prove to be confirmatory or otherwise with respect to the various ages of intrusion now held: (i) A study of the heavy mineral suite of the near-by sediments. (ii) A study of the antigoritic or non-antigoritic nature of a suitable intrusion, along with the grade of regional metamorphism (if any) in near-by rocks of suitable composition.

The Antigorites and the Formation of Antigorite

Antigoritic serpentinites have been found in the Cawarral-Tungamull area where the ultramafics are almost completely antigoritic, and in the southern portion of the Mary Valley belt. Under the microscope the habit of the antigorite varies from flaky to prismatic, often with the characteristic " thorn-structure " developed. Prisms are often arranged in sheaves and bundles, whilst the finer grained varieties are often " felted ". In the Queensland rocks the alteration of chrysotilic serpentine and bastite pseudomorphs to antigorite is similar to that described by Benson (1913, Part III, p. 674) from certain areas in New South Wales. Veinlets of cross-fibre chrysotile also reveal antigoritization.

Examination of the Queensland ultramafics revealed that the non-antigoritic rocks occurred in areas where regional metamorphism of the enclosing rocks is extremely slight or absent. In the Mary Valley the ultramafic rocks (5 miles in width) have been intruded by a complex of evenly distributed mafic sills, dykes, and pyroclasts. Later, both the serpentinites and mafic rocks were subjected to various grades of regional metamorphism. Over a north–south distance of 10 miles these metamorphosed mafic rocks (the Kandanga Metabasites) are now represented progressively by rocks of the greenschist and albite-epidote amphibolite facies (Turner, 1948).

Rocks from the greenschist facies of the Metabasites include chlorite (-epidote-sphene) schists, albite actinolite (-chlorite) schists and incompletely reconstituted epidiorites, with relict pyroxene and basic plagioclase feldspar, albite, actinolite or epidote, and chlorite. The metamorphic felspar in these rocks is an acid albite, averaging Ab_{95}.

In the albite-epidote amphibolite facies, the Kandanga Metabasites are represented by albite-amphibole schists, zoisite-sphene amphibolites and albite-epidote-sphene (-quartz) amphibolites. The metamorphic feldspar is a basic albite or acid oligoclase, whilst in addition to hornblende, amphiboles with deep blue-green tints parallel to Z are present. From a chemical standpoint, the pre-serpentinite representatives of the Brisbane Metamorphics in the area (cherts, jaspers, and thoroughly reworked argillites that suffered strong mechanical and chemical decay before final deposition) are poorly suited for the metamorphic regeneration of new minerals.

Sectioning of specimens taken parallel with, and at right angles to the serpentinite belt, indicates that the over-all areal distribution of antigorite is such that effective transformation of chrysotile serpentine to antigorite takes place where the associated Metabasites belong to the albite-epidote amphibolite facies. Antigoritization is believed to be dependent mainly on temperature. Shearing stresses are not thought to be important because of the preservation in the antigorites of such delicate pre-metamorphic structures as the network of magnetite granules and the parallel lines of iron ore, set free during serpentinization of olivine and orthopyroxene respectively. Evidence also indicates that the schistosity of the associated Metabasites is essentially crystalloblastic in origin and hence mainly a temperature effect. Pore solutions probably catalyse the formation of antigorite since the production of certain associated Metabasites, e.g. epidote-chlorite-calcite-sphene rocks, has occurred in the presence of such solutions. The increased temperatures necessary for the metamorphism of the Mary Valley rocks are best referred to magmatic injection at depth, since granitic rocks occur on both sides of the ultramafics, along with widespread satellitic porphyries and porphyrites in the belt proper.

Hess, Dengo, and Smith (1952, p. 74) placed the metamorphic grade at which antigoritization takes place as above the chlorite-biotite subfacies of the greenschist facies, but equal to or slightly less than the albite-epidote amphibolite facies.

The Serpentinite Contacts

In all specimens of undoubted country-rock at, or in the immediate vicinity of serpentinite contacts (including cherts, jaspers, mudstones, mica-phyllites), no evidence of thermal induration or hornfelsic structures has been noted. This feature of low-temperature intrusion of orogenic serpentinites is a common one (cf. Hess, 1938, p. 325). However, at various points along the serpentinite belt, the intruded rocks near the contacts frequently carry considerable amounts of obviously introduced calcite and quartz. At Pine Mountain, the contact cherts of the Neranleigh-Fernvale Group carry sufficient calcite to be termed siliceous limestones. The calcite frequently occurs as fissure-fillings in the brittle siliceous rocks, brecciated during the period of intrusion of the ultramafics. Sectioning of specimens at right angles to the contacts shows a progressive decrease in the carbonate content. The serpentinites themselves carry minor calcite but in no amount comparable with the country-rock.

On a tributary of Kingaham Creek in the Mary Valley, interesting phases occur where the serpentinite abuts against quartzose schists. The latter rocks at the contact contain up to 40 per cent biotite whilst the serpentinites are now represented by tremolite-actinolite rocks, with minor (?) gedrite and chlorite. Acid igneous rocks outcrop in the vicinity. The geological environment and rock associations are comparable with the higher temperature type of metamorphic differentiation occurring at serpentinite-siliceous country-rock contacts, described by Phillips and Hess (1936).

Where the ultramafics have been intruded by later igneous rocks (granites, porphyries, porphyrites, gabbros undergoing saussuritization, etc.), with consequent hydrothermal activity, metasomatic changes occur in the vicinity of the contacts, e.g. the Northbrook serpentinite has been highly altered to talc (-tremolite) schists.

Some Aspects of the Petrogenesis of the Serpentinites

Bowen and Tuttle (1949) have shown the absence of liquid phases at temperatures of up to 1,000° C. and 15,000 lb./in.^{2} in the system $MgO.SiO_2.H_2O$. Their experimental results preclude the possibility of a dunitic magma with large quantities of water (10 per cent or more), or a peridotitic magma having a composition equivalent to olivine plus orthopyroxene plus water and corresponding approximately or exactly to serpentine in composition, existing at temperatures less

than 1,000° C.—a temperature certainly not in accord with observations of metamorphism at serpentinite contacts. Bowen and Tuttle (op. cit.) conclude that ultramafics can be intruded only in the solid state and derive those liquids necessary for serpentinization from extraneous sources, e.g. from younger near-by acid granites or from the sediments into which intrusion occurred. In order to explain many world-wide characteristics of orogenic ultramafics, Hess (1938) postulated a primary peridotite magma, in which autoserpentinization of early formed olivine and orthopyroxene occurs in the presence of its own highly aqueous mother liquor. The high water content of this primary ultramafic magma would be supposed to lower its temperature sufficiently to account for the widespread evidence of low temperature of intrusion around orogenic serpentinites. In the discussion which follows, it is intended to do no more than compare these two conflicting hypotheses, as they stand in the light of data derived from examination of the Queensland serpentinites; accent will be placed on the mode of intrusion and serpentinization of the ultramafic rocks.

Examination of the Queensland ultramafics at widely distant localities has furnished the following information:—

(i) Irrespective of the size of the intrusion or the nature of the invaded rocks, all of the original olivine, together with the majority of the orthopyroxene has been serpentinized.[1] The degree of serpentinization of the orthopyroxene bears no relation to the margins of the intrusions. The water content of the serpentinites would average between 13 and 14 per cent.

(ii) Throughout the ultramafic belt there is a striking similarity, and indeed a monotony, in the mineralogy and petrography of the serpentinites. No feldspathic peridotitic types have been recognized. The post-serpentinite gabbros (usually extensively saussuritized) and dolerites are insignificant in bulk compared with the ultramafics.

Before acceptance of the origin of the serpentinites in the manner postulated by Bowen and Tuttle, it is thought that the following points should be carefully considered:—

(1) There does not appear to have been an adequate source of extra-ultramafic solutions in quantities sufficient for complete serpentinization to take place. Serpentinization was earlier than, and was not related to expulsions of hydrothermal solutions from later magmatic sources, since the serpentinite intrusions may be unassociated in the field with later igneous activity. Where this is present all stages of

[1] In the ultramafic rocks from both States, so far the writer has observed relict olivine only in one specimen. This occurs in a slice from a partially serpentinized dunite from the Nundle Area, N.S.W., in Professor W. N. Benson's collection in the Department of Mineralogy and Petrology, Cambridge.

replacement of serpentine by anthophyllite, tremolite, and talc occurs in the vicinity of the ultramafic contacts. The derivation of the serpentinizing fluids from the intruded sediments also presents difficulties. As an example, the Pine Mountain serpentinites, dominantly enclosed by cherts and jaspers of the Neranleigh-Fernvale Group, may be considered. Carefully selected specimens of chert and jasper showed a loss on ignition at 1,100° C. of 0·45 per cent to 2·25 per cent. The regular ignition loss is a reflection on any secondary material deposited along fracture planes in a series of specimens ranging from compact (the most common) to varieties brecciated as a result of intrusion of the ultramafics. These breccias contain normal fragments of chert. On available analysis an average optimum water content of 1–2 per cent could be assigned to the Neranleigh-Fernvale and most of the earlier representatives of the Brisbane Metamorphics. These then could not be regarded as " wet " rocks. An observation in the Barnard River Valley, New South Wales, by Osborne (1950, p. 67) " suggests that the serpentine had invaded rocks previously jasperized ". The same writer (op. cit., p. 68) also mentions several masses of jasper embedded in the serpentinite at Wood's Reef, New South Wales. It also would appear that induration of the country rocks had occurred prior to intrusion of the ultramafics and that the above figures are representative of conditions prevailing at the time of intrusion. It is difficult to reconcile complete serpentinization of large bodies of ultramafic with progressive desiccation of the country rock, especially if two phases of ultramafic activity occurred. Evidence indicates that such has been the case and that serpentinization of the first phase predated the intrusion of the second phase. The linear distribution of the intrusions along major fractures suggests that the movement to their present positions was dominantly vertical, with little or no lateral component, such vertical movement occurring through thousands of feet of Brisbane Metamorphics. These latter rocks are presumably underlain by a highly metamorphosed Pre-Cambrian basement, not a suitable contributory environment towards serpentinization (Hess, 1938, p. 331; Benson, 1918, p. 705). The larger intrusions of ultramafic rock can hardly be regarded as " pods ". In many instances their vertical or near-vertical schistosity parallels the walls of the fracture, and would indicate that the serpentinite bodies have deep-seated " roots ". Whether the vast quantities of serpentinizing fluids are referable to solutions at depth migrating along the fault plane is debatable, for mineralogical evidence of their former presence should be available in those areas of the fault zone where serpentinites do not occur. Such evidence has not been recognized.

(2) If intrusion of the ultramafic rocks occurred in the solid state,

even in the presence of small quantities of lubricating intergranular fluid, it is conceivable that, as the bodies moved up relatively narrow fissures to the accompaniment of strong overthrust faulting, the effects of this type of intrusion would be reflected in the presence in these rocks of intergranular movement and cataclasis, e.g. features present in the dunite-mylonites of St. Paul's Rocks, Atlantic, described by Tilley (1947). Yet the serpentinized dunites and harzburgites usually possess broad, well-defined, frequently equidimensional mesh-structure over large areas, with the attendant network of magnetite granules. Relict pyroxene and large bastite pseudomorphs up to several centimetres in length show no indication of cataclasis. Branching " dendritic " spinel is unsheared. The post-serpentinite gabbros (usually extensively saussuritized), occurring as relatively small bodies internally to the serpentinite intrusions, furnish a different picture. In these rocks, cataclastic effects are widespread. The pyroxenes show a wide variety of mechanical changes—shadow extinction, bending, fracture, and " mortar-structure ". Some of the finer grained types are granulitized. The final stages of consolidation of these rocks, whilst in a largely crystalline condition, took place under tectonic conditions very similar to the ultramafics. The whole impression gained from the serpentinized peridotites, even the inclusions, is one of crystallization and serpentinization in situ.

(3) The serpentinite belt furnishes several instances of phenomena suggestive of magmatic conditions. In this respect may be mentioned the myrmekite-like picotite which appears to form intergrowths with both olivine and pyroxene. It is difficult to reconcile the presence in the serpentinite of chromite lodes, e.g. in the Cawarral-Tungamull area, and the nickel-cobalt mineralization in the vicinity of Kilkivan, at the northern end of the Mary Valley belt, under conditions " where there is no reason to suppose that the crystalline mass initially has more than traces of water " (Bowen and Tuttle, op. cit., p. 455). Jones (1947, p. 36) states that the Queensland serpentinites were " responsible for quite extensive though not very rich mineralization by means of hydro-thermal solutions, in addition to the magmatic segregation of chromite ".

The derivation of the ultramafic rocks by any mechanism of differentiation from basaltic magma would have to be an exceedingly perfect one to fulfil many widely distributed features in the ultramafics. In particular, the monotony of petrography of the intrusions over considerable distances, the lack of gradational feldspathic peridotitic types and the absence or insignificant bulk of the later basic igneous rocks, should be stressed. The field and microscopic data on the serpentinites are more in accord with the hypothesis of Hess (1938), a concept incompatible with established experimental data.

Serpentinization of the ultramafics appears to have resulted from the activity of solutions from within rather than from without. The development of calcite and quartz along country-rock contacts in quantities greater than in the ultramafics is in agreement with the statement of Benson (1918*a*, p. 709) that " magmatic waters, highly charged with silica and carbonic acid, etc., have been emitted in connection with certain peridotitic intrusions ". It is believed that future intensive research on the serpentinites should be directed towards elucidation of many problems present in the schistose varieties. Why are they areally important and yet reveal little evidence of the derivation of the serpentine from pre-existing ferro-magnesian minerals? It might be suggested that the schistose rocks represent a mobile phase developed from, and subsequent to, serpentinization of the massive varieties. Any hypothesis along these lines would have to explain the various relationships present in the breccias and inclusions. One would expect to find every gradation existing between the two types. Large areas of the completely serpentinized massive phase occur adjacent to large areas of the schistose phase in the one intrusion, whilst the massive phase may terminate against country-rock contacts. The problem of serpentinite genesis has always been highly debatable. As in every problem, accumulation of fresh data may lead to new conclusions. It is hoped that certain interpretations in the present paper will perhaps serve to indicate new lines of research and stimulate new ideas.

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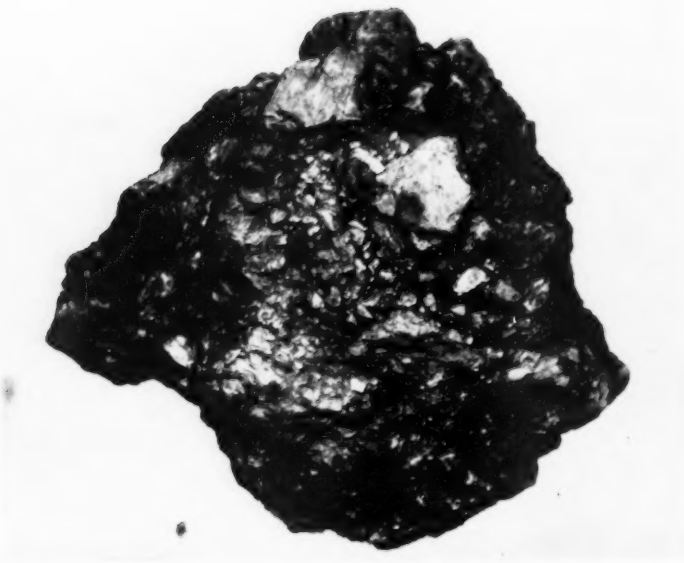

FIG. 1.

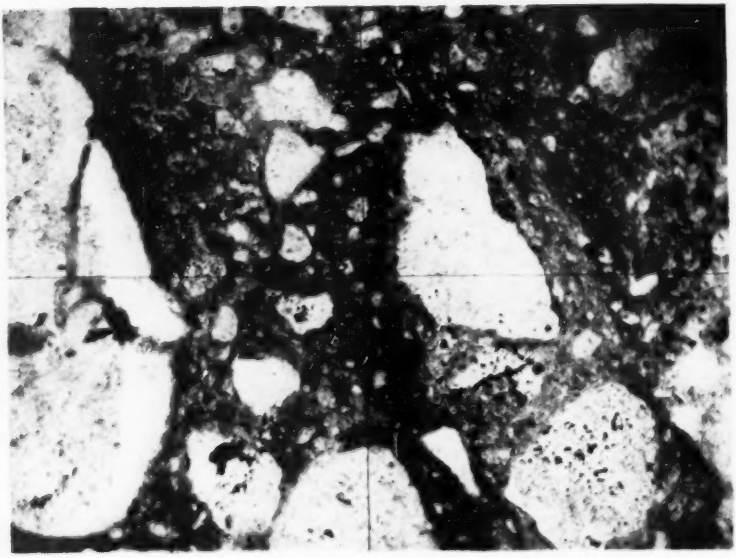

FIG. 2.

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EXPLANATION OF PLATE XII

FIG. 1.—Serpentinite-breccia showing size variation and angularity of fragments. (Approx. $\frac{2}{3}$ natural size.)

FIG. 2.—Photomicrograph of fine grained serpentinite-breccia; angular chips of mesh-serpentine are enclosed in a serpentinous matrix ($\times$ 60).

Cross-Bedding and Grain Size in the Lower Cretaceous Sands of East Anglia

By Walther Schwarzacher

Abstract

The cross-bedding of the Lower Cretaceous sands between Leighton Buzzard and Hunstanton has been studied. This paper describes new evidence which throws further light on the depositional environment and palaeogeography of these sediments.

Both cross-bedding and grain size distribution were investigated in detail in the Woburn Sands (Aptian). These beds lie between Cambridge and Leighton Buzzard (Cameron, in Kirkaldy, 1939, p. 407), and are the youngest " Lower Greensand " of East Anglia. Rather less attention has been paid to the pre-Barremian Sandringham Sands (Downham to Hunstanton). Unfortunately, the Carstone of Hunstanton shows no cross-bedding.

Dip of the Cross-bedding

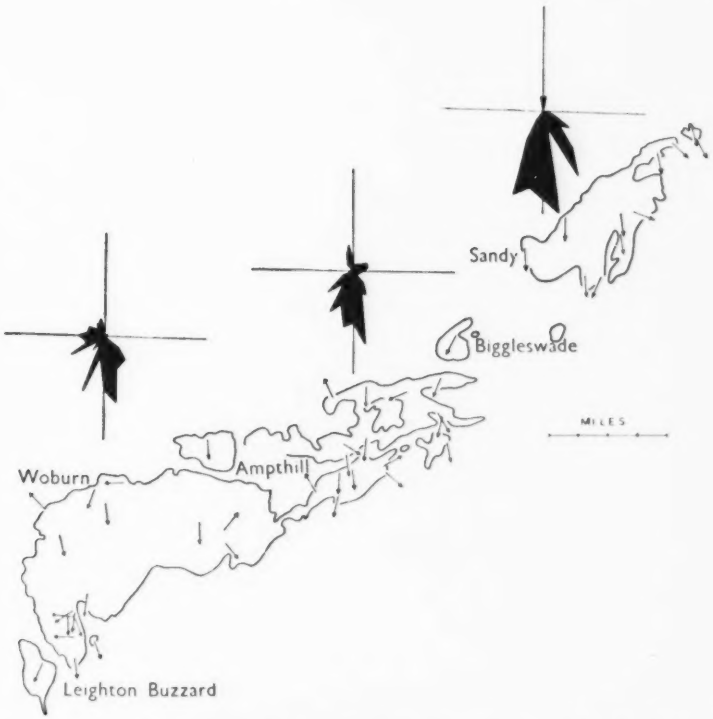
THE first point to be investigated was the possibility of any regularity in the dips of the cross-bedding throughout the area. Seven hundred measurements of dip-directions were taken, and are represented in Text-figs. 1 and 2. The Woburn Sands outcrop (Text-fig. 1) is divided into three sub-areas: that N.E. of Sandy being represented by seventy-seven cross-bedding units,[1] that between Ampthill and Biggleswade by 147 units, and that between Ampthill and Leighton Buzzard by 135 units. The dip given for each unit is an average of two or more readings. The Sandringham Sands (Text-fig. 2) were treated in two sub-areas, viz. from Hunstanton to Castle Rising with fifty-six units, and from Castle Rising to Downham with seventy-nine units.

The azimuthal distribution of dip for each sub-area is shown in the figures. In Text-fig. 1 the corresponding standard deviations are from S.W. to N.E. 44·6°, 42·6°, and 28·1°; in Text-fig. 2, they are 41° and 46° from S. to N. Owing to lack of statistical homogeneity in the sub-areas it is impossible to obtain the standard error of these quantities. Likewise, it is impossible to give for each exposure a full statistical description (cf. Reiche, 1938; Jüngst, 1938). They are therefore represented only by their arithmetic mean dip-directions. Very roughly, the standard deviation of the dip-directions per exposure (locality) is around 20°–30°.

Several localities, mostly in the Sandringham Sands, have cross-

[1] For statistical purposes a cross-bedding unit is defined as any thickness of sediment which shows cross lamination formed without any visible break in deposition.

bedding of the " herring-bone " type, east- and west-dipping beds alternating in a single section. Where less than 5 per cent of the exposed cross-bedding dipped in a particular direction, it was neglected. Where there was more than this, a double-pointed arrow is inserted on the maps to indicate the mean azimuth only, without implying any specific



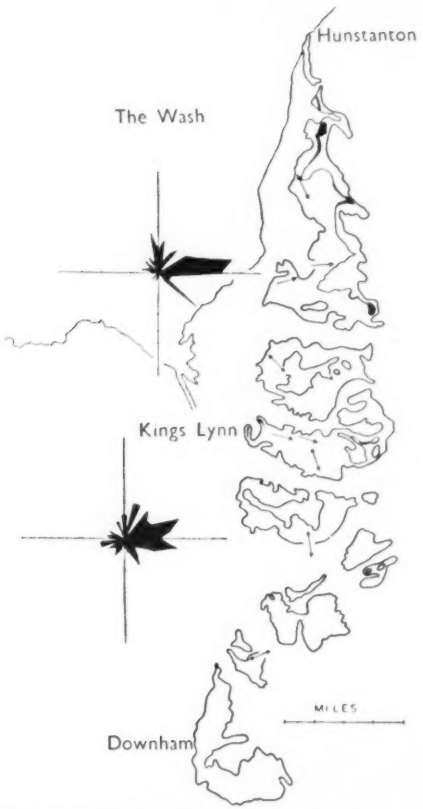
TEXT-FIG. 1.—The outcrop of the Woburn Sands. The mean dip direction of current bedding for each measured locality is shown by an arrow. The statistical distribution of dips for three sub-areas is shown in rose diagrams.

direction of dip. However, in the diagrams for the five sub-areas all the measurements have been included. Thus it can be seen that in Text-fig. 1 north-dipping cross-laminations are almost absent, and in Text-fig. 2 the westerly components are decidedly smaller than the easterly ones.

The mean dip (from all measurements) was found to be 29°, and any statistically significant variation in the amount of dip was less than 5°. The tectonic disturbance, however, is of the same order.

Structure of the Cross-bedding

The unit of cross-lamination in the Woburn Sands is usually tabular in shape (Shrock, 1948). In the Sandringham Sand the smaller units (5–20 cm. thick) often lie between parallel beds; the larger units (1–5 m.) sometimes have truncated tops showing no signs of topset beds.

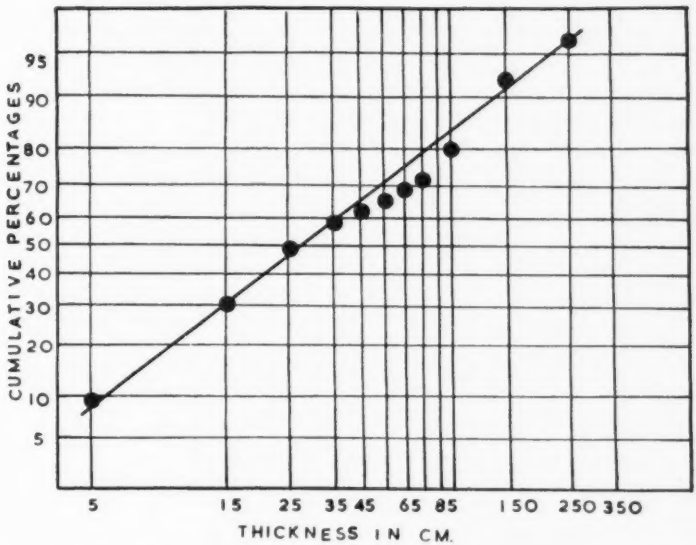


Text-fig. 2.—The outcrop of the Lower Cretaceous Sands in Norfolk. The Snettisham Clay is marked in black. The distribution of current bedding dips in the Sandringham Sands is indicated as in Text-fig 1.

The thickness-frequency curve for the Lower Greensand (100 randomly selected cross-bedding units from the Ampthill–Biggleswade area) conforms to a lognormal probability distribution (Text-fig. 3). The coefficient of variation of the log thickness—54 per cent—is large in comparison with the variation shown by some Pleistocene varves (e.g. 14 per cent, Pettijohn, 1949), and certain Triassic limestone—

dolomite varves (e.g. 29 per cent, Schwarzacher, 1947). The relation between median grain size and the thickness of a bed, as described by Fiege (1937) and Kurk (after Pettijohn, 1949, p. 141), also applies to this Lower Greensand area (Text-fig. 4).

Large- and small-scale cross-bedding are not homogeneously distributed throughout the area. Therefore, the thickness of the cross-laminated beds in each locality, and/or their median grain size, reflect distinct depositional conditions. The factors influencing the



TEXT-FIG. 3.—The cumulative curve of logarithmic thicknesses of 100 randomly selected cross-bedding units drawn on normal probability paper. The graph shows that such frequency curves conform to a lognormal distribution.

thickness of a cross-laminated unit are complex. Experimentally it can be shown that three factors are of primary importance: the relative height of the water level, the configuration of the bottom, and the amount of sediment supplied. To obtain cross-bedding artificially, the competence of the water-current has always to be adjusted to the *preceding* conditions. Consider for example an experimental channel through which a constant discharge of water flows. The competence of the current varies with the depth of water. Automatically, then, if coarse sand is used it will form thick cross-bedding units, and fine sand will form thinner ones. Now if the supply of sediment is stopped and the velocity is made to fluctuate, the sand will become more evenly

distributed and the cross-bedding units will become thinner. Again a plentiful supply of coarse sand will cause thick units to form.

It seems therefore reasonable to assume that cross-bedding units will thicken towards the source of supply of the sediment.

Mechanical Analysis of the Sand

Exposures were not sufficiently numerous to make a detailed study of the thickness of the cross-bedded units and their grain sizes. However, certain striking regularities in the stratigraphical distribution of the cross-laminated beds and their median grain size came to light. These have been summarized in two vertical sections, Text-fig. 5*a* (from near Clophill, E. of Ampthill), and 5*b* (from near Woburn), the datum line in each case being taken at the top of the Jurassic, and exposures lying as much as 3 miles from the line of profile being projected on to the sections in order to get a reasonable number of points. The notorious lateral variation of the sands implies that the two sections give at best only a very schematic idea of the distribution of particle size in the area. The data from which the sections were constructed follow in Tables I and II, being the results of mechanical analysis in a 7 ft. sedimentation tube, as described by Doeglas (1946).

Table 1 (Clophill)

National grid reference.	*Median grain size in μ.*	*Standard error* of median in μ.*	*Number of Samples.*
Under the 350 line.			
50802380	200	81	10
51182401	250	67	6
50822373	210	10	8
50992371	100	—	3
Under the 500 line.			
50722407	360	91	14
50892410	360	10	4
50752383	360	46	8
51222241	340	—	3
Above the 500 line.			
5092416	1,090	—	2
50802380	600	84	6
51182401	960	150	6
50992371	1,070	122	6
50782358	800	217	8
Above the 350 line.			
50632358	390	75	6
			90

TABLE 2 (WOBURN)

National grid reference.	*Median grain size in μ.*	*Standard error* of median in μ.*	*Number of Samples.*
Under the 250 line from NW to SW:			
49542361	220	17	8
49432350	200	0	6
50102330	230	64	12
Under the 350 line			
49402364	280	83	12
49252348	280	87	10
49922382	330	57	8
49402322	260	—	2
49252282	340	71	14
Under the 500 line.			
49122345	350	56	8
49192333	380	89	8
Above the 500 line.			
49352298	550	179	12
49972341	610	—	2
49372274	770	203	10
			112

* This is the error which has been found between different samples from the same locality.

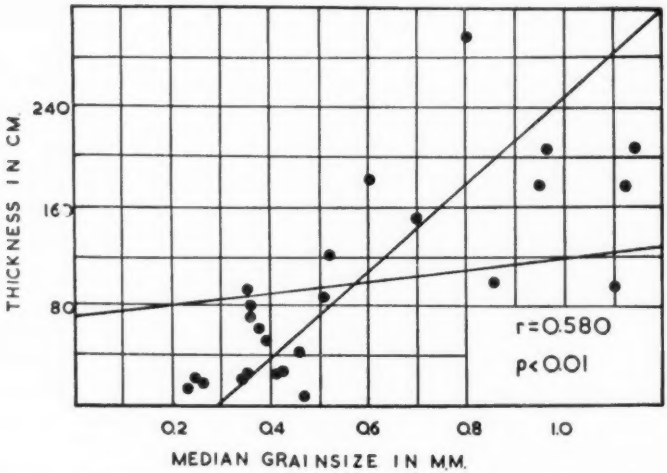
Around Clophill, thick (2–3 m.) coarse-grained cross-bedding units rest on an eroded surface of thin (5–20 cm.), fine-grained units. The latter have well-developed top- and bottom-set beds, dipping north. The eroded boundary can be mapped over several miles. Certain exposures in the fine beds have no cross-lamination, but only horizontal bedding, clearly shown by thin bands of clay. At one locality near Clophill, the clay reaches 6 feet in thickness.

Between Ampthill and Leighton Buzzard, as at Clophill, the Greensands show an increase in coarseness from the bottom upwards and a thickening of the coarse beds northwards. The source of the coarser sediments therefore lay towards the north and probably the finer facies were developed by winnowing from these.

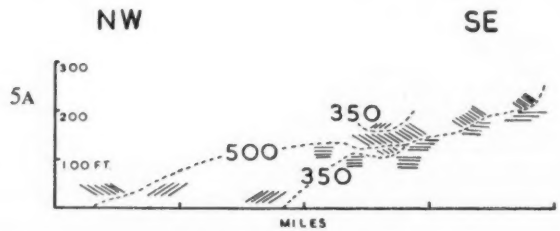
DEPOSITIONAL ENVIRONMENT

From the palaeogeographical point of view it is important to know whether the cross-bedding represents longshore drift, wave action, and/or delta formation, and hence to ascertain its relation to the coastline.

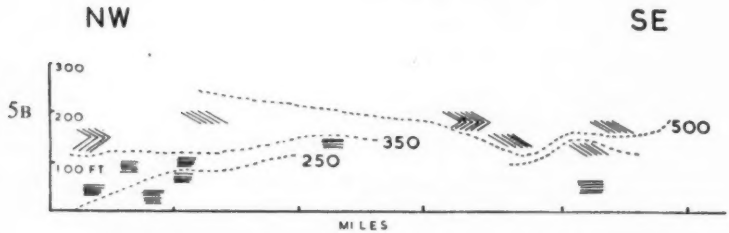
The evidence concerning modern sediments is meagre, but it seems doubtful if longshore drift could ever produce cross-lamination dipping



TEXT-FIG. 4.—Correlation diagram of median grainsizes (measured in mm.) and thicknesses of cross-bedding units (measured in cm.) from the Woburn Sands. The significant coefficient of correlation in the 25 examined exposures is r = 0·580.



TEXT-FIG. 5—(*a*) Diagrammatic section through the Woburn Sands near Clophill. Horizontal shading indicates exposures with small scale cross-bedding, inclined shading, exposures with large scale cross-bedding. Broken lines connect places with 350μ, 500μ, median grainsize. For data see table I (p. 326).



(*b*) Diagrammatic section through the Woburn Sands near Woburn. Broken lines connect places with 250μ, 350μ, 500μ, median grain size. For data see table II (p. 327).

consistently parallel to a coast. Observations on shores having artificial structures show that longshore drift causes the beach to grow seawards (Sheppard, 1948). I do not know of any work on the structure of such beaches, but I should expect their cross-bedding (if any) to dip predominantly seawards. Examination of certain elongated sand-bars farther offshore (Jüngst, 1938) has so far revealed no structures of this sort. They were, indeed, too thin and streamlined to develop cross-lamination. Moreover, no undoubted case of current-bedding due to longshore drift in ancient sediments has yet been reported. Cross-laminations dipping perpendicularly to the coast (mainly off-shore, but also dipping towards the shore) have been detected in modern beaches (Thompson, 1937), as well as in ancient off-shore sediments (Jüngst, 1938; McKee, 1940).

However, in the Woburn Sands, the large-scale cross-bedding must surely indicate submarine delta-formation, and the direction of thinning of the units an increasing distance from the shore. The contiguous beds in the Sandringham Sand with opposing dips may then suggest a more or less horizontal floor in the centre of the basin. Only a few doubtful ripple-marks occur, and no extreme channelling was seen.

Palaeogeography

The cross-bedding in the Woburn Sands (Text-fig. 1) indicates that in Aptian times a shore-line lay to the north. It ran more nearly E.–W. than either of those suggested by Kirkaldy (1939, fig. 37, p. 412). No influence of the London platform was discernible. The data are too few to show a reflection of Rastall's N.W.–S.E. anticlines (1925) in the distribution of dips. Possibly a tongue of coarse sand near Leighton Buzzard (also observed by Lamplugh, 1922, p. 56) is related to a small anticline N.W. of Dunstable.

Earlier, during Sandringham Sands times, a N.–S. coastline had apparently lain near the Wash, between Norfolk and Lincolnshire (cf. Kirkaldy, 1939). Again no influence of the London platform was discernible.

The experimental work was carried out at the Sedgwick Museum, Cambridge, with the aid of a British Council scholarship. Professor P. Allan took much trouble over reading and revising this paper and Mr. R. G. C. Bathurst also helped to correct the manuscript.

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DEPARTMENT OF GEOLOGY,
QUEEN'S UNIVERSITY,
BELFAST.

ANNOUNCEMENT

University of London : Special Lectures

The following Special Lectures have been announced :—

Professor Leif Størmer (Oslo) : Trilobites and Merostomes, an old arthropod stock ; and Recent Studies in the Middle Ordovician of the Oslo Region. *December* 8th, 9th, and 11th, 1953, at King's College, Strand. The Chair at the first lecture will be taken by Dr. C. J. Stubblefield.

Professor R. A. Stirton (University of California) : Australian Mammalian Fauna and its Origin. *December* 1st, 1953, at University College, Gower Street. The Chair will be taken by Professor D. M. S. Watson.

Professor R. Tomaschek (Anglo-Iranian Oil Research Centre) : The Tides of the Solid Earth. *November* 27th and 30th, 1953, at University College, Gower Street. The Chair at the first lecture will be taken by Professor L. Hawkes.

Admission to all is free, without ticket.

Bajocian Ammonites Collected by Sir Henry Hayden near Kampadzong, Tibet

By W. J. Arkell

(PLATES XIII and XIV)

Abstract

Seven species of ammonites from the Lungma Limestone of Tibet are described and figured. The assemblage is Middle Bajocian, of the Sowerbyi and Sauzei Zones, and possible also Humphriesianum Zone, and indicates that the formation is condensed. Its affinities are chiefly with European faunas, but it has one species in common with the Pamirs and another with Western Australia.

I. Introduction

THROUGH the kindness of the Director of the Geological Survey of India, Dr. M. S. Krishnan, I was loaned in 1953 a small collection of ammonites brought by Sir Henry Hayden in 1903 from the Lungma or Langma Limestone of Mekyigunru, 8 miles south of Kampadzong, Tibet. They were recorded (Hayden, 1907, p. 33) as follows: " *Harpoceras* (*Dorsetensia*) *complanatum* Buckman, *Harpoceras* several other species (these will probably be found to include other genera, such as *Sonninia*), *Stepheoceras* cf. *humphriesianum* Sow.," together with various brachiopods and lamellibranchs. The fauna was correctly dated to the English Inferior Oolite.

The Lungma Limestone was described as about 50 feet thick. It divides thick slates, quartzite, and conglomerate assigned to the Lias below, from black and grey shale and quartzite above; overlying these are the Spiti Shales. Photographs of the fossiliferous outcrops are shown in Hayden's Plate 9. No ammonites were found except in the Lungma Limestone and Spiti Shales.

The matrix of the Bajocian specimens is a hard, grey and brown, mottled, shelly limestone, patchily speckled with black ooliths. The shells of lamellibranchs and ammonites are preserved, but the ammonite specimens are mostly casts, the test having broken away during extraction in the field.

This occurrence is of special interest owing to its extreme isolation from all other known occurrences of marine Bajocian. The nearest recorded Bajocian ammonites are crushed and distorted specimens collected in the south-eastern Pamirs (Vyalov, 1935), and recorded as *Oppelia* cf. *subradiata* (Sow.), *Lissoceras* cf. *oolithicum* (d'Orb.), *Emileia* cf. *polyschides* (Waagen), *Dorsetensia* cf. *regrediens* (Haug), ? *Ermoceras* sp., and *Lytoceras* sp. This is about 1,100 miles from Kampadzong. The better-known Persian occurrences, near Ravar and in the Elburz, are about 2,000 miles, and Jebel Tuwaiq in Central Arabia about 2,600 miles distant. To the N.E. the marine Bajocian

in Eastern Transbaikal is about 2,000 miles away, and that to the S.E. in the Moluccas about 3,400 miles.

Thus the Tibetan occurrence is the only known link (besides the little-known one in the Pamirs) between the Bajocian occurrences of East and West Asia, otherwise separated by 3,000 miles (Persia–Siberia), or 5,000 miles (Persia–Indonesia). Since Hayden was not an ammonite specialist, and the collections do not seem to have been examined for nearly fifty years, it seemed desirable to subject at least the ammonites to a fresh scrutiny in the light of more modern ideas of systematics and correlation. I am therefore extremely grateful to Dr. Krishnan for acceding to my request to have the collection on loan for study.

I am indebted also to Mr. A. G. Brighton for help with Buckman's collection and types in the Sedgwick Museum, and to Mr. R. V. Melville for sending me on loan Buckman's other relevant types from the Geological Survey Museum.

II. Description of the Species

Sonninia aff. *dominans* S. Buckman (Plate XIV, figs. 8*a*, *b*)

Cf. *Sonninia dominans* S. Buckman, 1892–4, p. 435, pl. lxvii, figs. 1, 2; pl. xciv, fig. 1.

A small specimen, 44 mm. in diameter, belongs to this common group, but specific determination cannot be definite. The tubercles are large and die out at about the same diameter as in *S. dominans*, and the ribbing thereafter is of about the same strength and tends to be twinned to about the same extent. But unlike all or nearly all of the English species, the nucleus before a diameter of about 10–12 mm., though strongly ribbed, is not tuberculate. No reliance can be placed upon the ornamentation of nuclei at such small diameters as guides in taxonomy (K9/247).

Witchellia tibetica sp. nov. (Plate XIV, figs. 7*a*, *b*)

The holotype is 48 mm. in diameter and slightly distorted, but is quite distinctive. The ribbing is fine, irregularly twinned, with occasional flared and tuberculate ribs up to a diameter of 25 mm. The venter is tabulate and tricarinate-bisulcate, all three keels reaching the same height.

The nearest figured species is *Witchellia glauca* Buckman (1925, vol. vi, pl. dxciv) from the upper Sowerbyi Zone of Dorset, but *W. tibetica* is more evolute and more finely ribbed. Its evolute coiling is a departure from the type species, *W. laeviuscula* Sowerby sp. (refigured Buckman, 1927, vol. vii, pl. dccxlv), and brings it closer to "*Zugophorites*" *zugophorus* Buckman (1922, vol. iv, pl. cccxli), and "*Gelasinites*" *gelasinus* Buckman (1925, vol. vi, pl. dxciii), both from the

middle subzone of the Sowerbyi Zone and both included in *Witchellia* by Spath (1936, p. 5) (K9/245; ? 240).

WITCHELLIA aff. *PLATYMORPHA* S. Buckman (Plate XIV, figs. 6*a*, *b*)

Cf. *Witchellia platymorpha* Buckman, 1925, vol. vi, pl. dlxxx, A and B.

A cast 124 mm. in diameter with half a whorl of body-chamber, and another about 75 mm., wholly septate and slightly distorted, belong to a new species which will be described and figured on the basis of better material from Western Australia. It belongs to the group of *W. platymorpha*, characterized by wide, distant, feeble ribbing, lost early, and a high flat-sided whorl with tabulate, strongly tricarinate venter. Closely allied are *W. glauca* Buckman (1925, vol. vj, pl. dxciv), which forms a link with *W. tibetica*, and *W.* (" *Dundryites* ") *pavimentaria* Buckman, sp. (1927, vol. vii, pl. dccli), which occurs in Eastern Persia (Spath, 1936, p. 6, pl. i, fig. 2); but that is more involute and more regularly, closely, and sharply ribbed.

Witchellia furticarinata as figured by Haug (1893, pl. viii, fig. 3 *a–c* only) is also close, but Haug's larger specimen (fig. 4) has a rounded and unicarinate venter, and according to Dorn's drawings (1935, p. 49, text-fig. pl. iv, fig. 8) the whorl-section of Quenstedt's *furticarinata* is that of a *Sonninia* (K 9/231, 248; ? 235).

DORSETENSIA cf. *ROMANOIDES* (H. Douvillé) (Plate XIII, figs. 1*a*, *b*)

Cf. *Ludwigia romanoides* H. Douvillé, 1885, p. 28, pl. iii, figs. 3–5.

A well-preserved specimen with test, 66 mm. in diameter, had been identified previously with *Dorsetensia pulchra* Buckman (1891, pl. lii, figs. 25–7). But although the resemblance is considerable, comparison with the type-specimen (in the Sedgwick Museum) reveals that the Tibetan specimen has a less compressed whorl-section, and in that respect it agrees perfectly in ventral view with *D. romanoides* (H. Douvillé). Although the umbilical whorls (up to a diameter of about 27 mm.) are more distinctly ribbed than in *D. romanoides*, this is probably no more than individual variation. In Dorn's figure (1935, pl. xx, fig. 3: Sauzei Zone, Franconia) the ribs are wider and more distant (K9/237).

DORSETENSIA cf. *REGREDIENS* (Haug) (Plate XIII, figs. 2*a*, *b*, 3)

Cf. *Dorsetensia edouardiana* Buckman, 1892, p. 304, pl. lii, figs. 8–24 (*non* d'Orbigny, *teste* Haug).

Cf. *Witchellia regrediens* Haug, 1893, p. 318, pl. x, fig. 7.

According to Haug (1893, p. 319) the species figured as *D. edouardiana* by Buckman, which Haug renamed *regrediens*, differs from *D. edouardiana* d'Orbigny (which Haug equated with *D. pulchra* Buckman) by being more evolute, by lacking verticality in the umbilical wall

and by having a bifid first lateral lobe. Two Tibetan specimens 28 mm. and 26 mm. in diameter show no sutures, but otherwise agree well with Buckman's and Haug's figures, except that the ribbing is slightly more inflected, as in Buckman's figures rather than Haug's (K9/232, 241, 243).

This species is also tentatively recorded from the Pamirs (Vyalov, 1935).

DORSETENSIA ? HAYDENI sp. nov. (Plate XIII, figs. 5*a*, *b*, *c*; Plate XIV, figs. 9*a*, *b*)

Three well-preserved specimens, of which the two largest are about 53 mm. and 62 mm. in diameter and have the body-chamber with peristome. There is a very shallow lateral lappet, hardly more than a sinuosity, but a considerable ventral lappet, as in *Dorsetensia complanata* Buckman (1892, pl. liii, figs. 1, 2). The shell is compressed, evolute, with gentle umbilical slope, feebly ribbed, the ribs distant, irregular, confined to the middle of the whorl sides. The keel is blunt, hollow but not floored, bounded by incipient grooves. Sutures with small blunt lobes, the first lateral trifid.

Dorsetensia complanata Buckman has the same apertural features and similar rather irregular and distant rubbing, but it is more evolute and larger and the venter shows no signs of incipient grooves. (There are numerous specimens in the Sedgwick Museum, labelled by Buckman.)

"*Sonninia*" *deltafalcata* Haug (1893, p. 293, pl. ix, fig. 9) is a closer match in side view, but is decidedly less compressed and also larger and less feebly ribbed. Until the lectotype (Quenstedt, 1858, pl. 53, fig. 8, designated by Dorn, 1935, p. 114) is refigured, it remains doubtful whether Haug correctly identified his specimens, for Quenstedt's figure shows a smaller umbilicus, and there is no ventral view. There is certainly not specific identity with Borissjak's Donetz species confused with *deltafurcata* by Hiltermann (1939, p. 170).

In their gentle umbilical slope these species come close to the Australian *Dorsetensia clarkei* Crick (see Spath, 1939, p. 129, pl. ii, figs. 1, 2), which there is reason to believe should be transferred to *Fontannesia* (I have recently studied about 100 specimens of this and allied species from Western Australia, collected by Mr. P. E. Playford) (K 9/239, 244, 246).

EMILEIA (*FROGDENITES*) sp. (Plate XIII, figs. 4*a*, *b*, *c*)

The only stephanoceratid sent me is only 18 mm. in diameter but can be identified as belonging to the dwarf genus *Frogdenites* Buckman (1921, vol. iii, pl. ccxv, and 1923, vol. v, pl. cdxxx), of the English Sauzei Zone. This is less surprising since a *Frogdenites*

has been figured from the Lower Yakoun formation of Queen Charlotte Islands, Canada (McLearn, 1929, p. 18, pl. xii, fig. 6), and another from the " Sauzei Limestones " of Peru (" *Sphaeroceras* " *broggianum* Lisson, 1937, pl. i). According to the label, this specimen was the basis for Hayden's record of *Stephanoceras humphriesianum*.

Frogdenites may be considered only a subgenus of *Emileia*, with which it is closely allied and contemporary. The Tibetan specimen is regarded as a *Frogdenites* rather than a nucleus of *Emileia* s.s. because, even though a cast, its ribbing is sharper, more wiry, and it already shows signs of sharp tubercles which on the test were probably spines (K 9/249).

III. The Age of the Fauna

All the ammonites seen from the Lungma Limestone are Middle Bajocian. The age-range is therefore not so great as that covered by the ammonites recorded from the Pamirs (Middle and Upper Bajocian), or Persia (Lower, Middle, and Upper), or Arabia (Middle and Upper), or Indonesia (Middle and Upper), and there is no comparison with Siberia (Lower and Upper, but so far no Middle). In Western Australia, near Geraldton, however, there is a limestone formation (Newmarracarra Limestone, maximum thickness 38 feet) containing a wholly Middle Bajocian fauna; Lower and Upper Bajocian are unknown in Australia. But although the Lungma and Newmarracarra Limestones are both undoubtedly Middle Bajocian, there is only one species in common, and the dominant Australian stephanoceratid genus, *Pseudotoites*, is not represented in the collection from Tibet.

The principal links are the Sonninidae (*Sonninia*, *Witchellia*, *Dorsetensia*, *Fontannesia*), some of which occur in all the outcrops—Arabia, Persia, Pamirs, Tibet, and Australia—except Indonesia, which is linked with Australia by stephanoceratids.

The species to which the Tibetan ammonites bear the closest comparison date them in the European succession as follows:—

Sonninia cf. *dominans* Buckman	Sowerbyi Zone
Witchellia tibetica nov.	" "
" cf. *platymorpha* Buckman	" "
Dorsetensia cf. *romanoides* (Douvillé)	Sowerbyi and Sauzei Zones
" cf. *regrediens* (Haug)	Humphriesianum Zone
" *haydeni* nov.	" "
Frogdenites sp.	Sauzei Zone.

Thus all three of Oppel's European zones of the Middle Bajocian may be represented in the Lungma Limestone. The ironshot lithology and high concentration of ammonites and other shells also proclaim a condensed deposit. Just how condensed such a 50 ft. limestone may be is suggested by the fact that in Oregon the Middle Bajocian

is 1,300 metres thick, while in the Caucasus the thickness of the Lower Bajocian is 2,000 metres.

Representation of the Humphriesianum Zone in the Lungma Limestone is very doubtful in the absence of characteristic Stephanoceratidae. It can only be said with certainty that the Sowerbyi and Sauzei Zones are represented.

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EXPLANATION OF PLATES

PLATE XIII

(Geological Survey of India registration numbers)

1*a*, 1*b*. *Dorsetensia* cf. *romanoides* (H. Douvillé). K9/237.
2*a*, *b*, 3. *Dorsetensia* cf. *regrediens* (Haug). K9/243, 232.
4*a*, *b*, *c*. *Emileia* (*Frogdenites*) sp. K9/249.
5*a*, *b*, *c*. *Dorsetensia haydeni* sp. nov. Holotype. K9/246.

All photos natural size, by A. Barlow.

PLATE XIV

(Geological Survey of India registration numbers)

6*a*, *b*. *Witchellia* aff. *platymorpha* Buckman. K9/248.
7*a*, *b*. *Witchellia tibetica* sp. nov. Holotype. K9/245.
8*a*, *b*. *Sonnonia* aff. *dominans* Buckman. K9/247.
9*a*, *b*. *Dorsetensia haydeni* sp. nov., showing peristome. K9/239.

All photos natural size, by A. Barlow.

1a

1b

4a

4b

4c

2a

5c

3

2b

5a

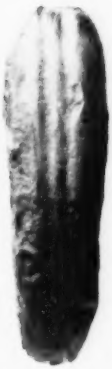
5b

BAJOCIAN AMMONITES FROM TIBET.

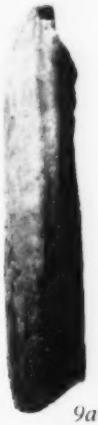
6a

6b

7a



7b



9a

8a

8b

9b

BAJOCIAN AMMONITES FROM TIBET.

Modal and Chemical Analyses in Regional Studies

By E. H. Timothy Whitten

Abstract

A technique for roughly evaluating the chemical variability within a suite of rocks is described and discussed in relation to the composition of the granite and quartzite of N.W. Donegal. The accuracy of this method is compared with more expensive chemical analysis.

During recent months considerable attention has been focused upon the degree of accuracy which may be expected in chemical analyses of silicate rocks (see, for example, Fairbairn *et. al.*, 1951, and Hamilton, 1952). In particular Fairbairn seems to indicate that frequently too great an importance has been attached to minor variations in chemical analyses (having varying degrees of accuracy). Because of the time and expense involved in chemical analysis the author investigated the practicability of estimating chemical variation from modal data. It was found that in examining the contact relations between Caledonian granite and Dalradian quartzite at the Bloody Foreland (north-west Donegal), and in studying the variability of this granite, results of remarkable consistency were obtained by calculating the chemical composition of the rocks from their modes. Such a technique has obvious limitations but it appears that the main chemical variations may sometimes be usefully demonstrated in this manner. It is hoped that this paper may stimulate the preparation of a greater number of modes, and hence lead to an examination of the usefulness of this approach in other areas and problems.

Technique Employed

The modal composition of each specimen was determined with the aid of a Dollar integrating micrometer. Such modes were always calculated from thin-sections having at least the maximum area which can be measured by the micrometer, and successive traverses were made 2 mm. apart. Published accounts usually do not state the size of the counts from which modes were calculated. Walker and Poldervaart (1949, p. 629) indicate that in studying the Karroo dolerites with a Dollar integrating micrometer " . . . the lengths of the traverses were over 200 times greater than the longest continuous intercept made by any mineral in the sections measured. Hence, the probable error is 1–2 per cent . . ." In nearly all cases their modes are given to one decimal place. Again, Heald (1950) in describing the Lovewell Mountain Quadrangle quotes the total number of slides modalized and averaged to give the composition of a variable rock mass, but seems to rely on single slides for the calculation of modes of particular rocks.

The standard textbook chemical compositions for the various phases have been employed for the determination of the weight of oxides present. As an example the case of microcline ($K_2O.Al_2O_3.6SiO_2$) is given. The formulae weights of the component oxides are 94·20, 101·94, and 60·1 for K_2O, Al_2O_3, and SiO_2 respectively. Hence the sum of these formulae weights for microcline is 556·74, whilst its specific gravity may be taken as 2·5. If the modal percentage of mineral in the rock is *d*, then the microcline present contains:—

$$\frac{101{\cdot}94 \times 2{\cdot}5 \times d}{556{\cdot}74} \text{ grams} = 0{\cdot}4575\ d \text{ grams of } Al_2O_3$$

$$\frac{94{\cdot}20 \times 2{\cdot}5 \times d}{556{\cdot}74} \text{ grams} = 0{\cdot}4231\ d \text{ grams of } K_2O$$

$$\frac{6 \times 60{\cdot}1 \times 2{\cdot}5 \times d}{556{\cdot}74} \text{ grams} = 1{\cdot}619\ d \text{ grams of } SiO_2$$

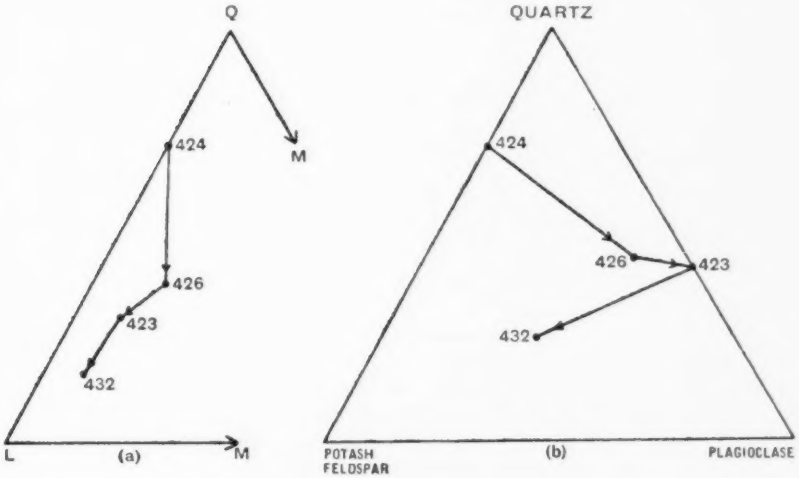
Values of this type were obtained for each mineral phase present in the rock, and the amounts of each oxide in the whole rock were then summed, and expressed as percentages—giving the weight per cent of oxides in the rock. These values, divided by the molecular weight of the oxide concerned, and expressed as percentages represent the molecular proportions per cent, from which von Wolff values may be directly determined and plotted on a ternary diagram (see Johannsen, 1931, pp. 113–6).

Since the results discussed in this paper have all been based on the study of coarse and medium grain-size rocks the "superposition error" involved in micrometric analysis (Elliott, 1952) is not of sufficient magnitude to be significant. So as to avoid emphasizing minor variability the values quoted, after reduction to percentages, have been corrected to two significant figures for values $>$ 1 per cent, and to one significant figure for values $<$ 1 per cent.

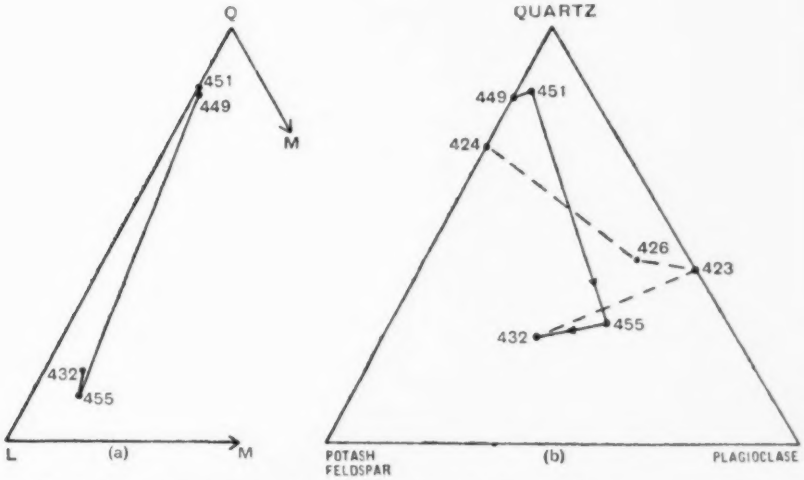
The calculation of von Wolff values from a prepared mode (per cent volume) involves tedious arithmetic, which takes 2½–3 hours per specimen. In addition the mode takes at least 1½ hours to complete. However, results obtained in this manner are obviously far less costly and time consuming than actual chemical analyses, and it is considered that provided certain safeguards are applied the results are often equally valuable.

Illustrative Examples from Bloody Foreland and Gweedore

1. The country rock at Bloody Foreland comprises massive and well-bedded Dalradian quartz-schist, and the contact between the granite and this rock is very irregular, transgressive, and steep. The granitic rock of the area is a biotite-hornblende quartz-rich adamellite (sp. 432 of Table I). Although the quartzite is marginally recrystallized and



TEXT-FIG. 1.—Variation at contact A, Bloody Foreland.
(*a*) von Wolff diagram, (*b*) partial mode diagram.
424 quartzite few inches from granite contact A.
426 quartzite ½ inch from granite contact A.
423 granite 2–3 inches from quartzite contact A.
432 regional granite at Bloody Foreland.



TEXT-FIG. 2.—Variation at contact B, Bloody Foreland.
(*a*) von Wolff diagram, (*b*) partial mode diagram.
449 quartzite 40–50 feet from granite contact B.
451 quartzite 10 feet from granite contact B.
455 coarse granite near quartzite contact B.
(Remaining numbers as Text-fig. 1).

enriched in feldspar, sharp contacts prevail and granite tongues occur along bedding planes.

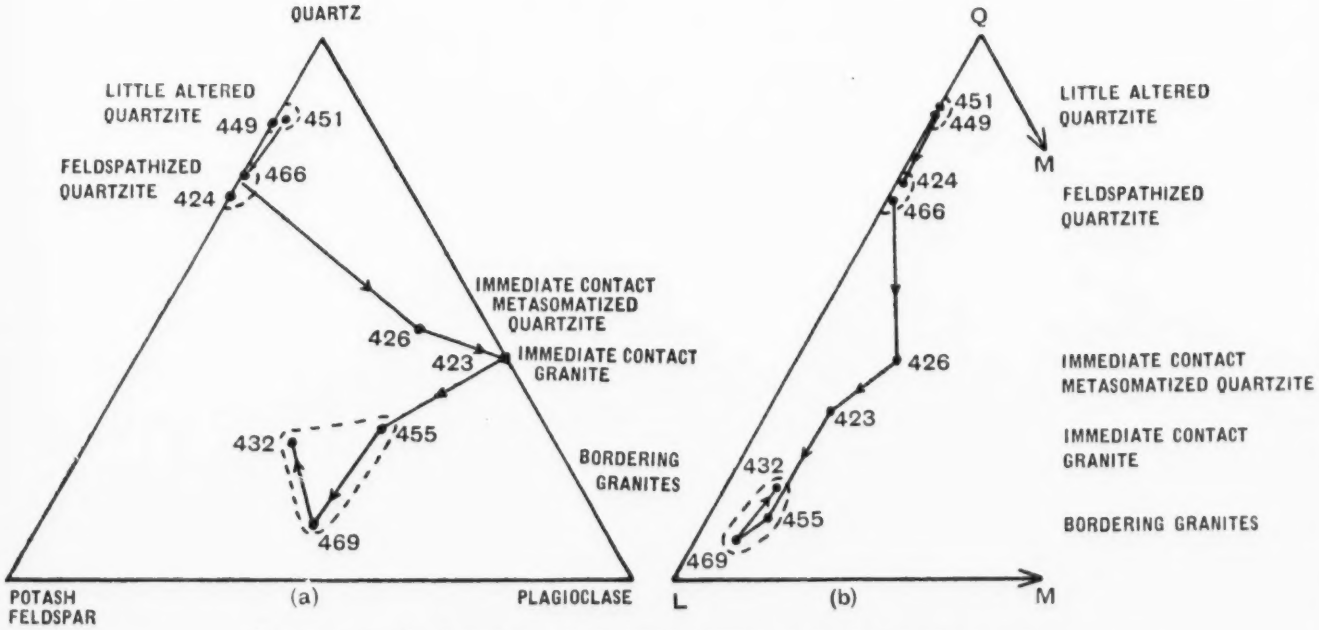
Table I records the modal variation occurring in specimens collected astride three of the exposed contacts, namely A, B, and C. Variation across contact A is indicated in Text-fig. 1 (*a*) and (*b*). The latter emphasizes the culmination of plagioclase in the contact zone. The former illustration shows a progressive basification, and a mafic culmination at the actual contact, in passing from the quartzite (sp. 424) a few inches from the granite to granite (sp. 423) 2–3 inches from the contact. Since these specimens have all been collected within inches of the contact, it is instructive to compare Text-fig. 1 with Text-fig. 2, since the latter only show progressive basification without any marked culminations. The specimens from which Text-fig. 2 has been constructed were collected several feet (rather than inches) from the contact. Detailed investigation in the field and laboratory suggests that the specimens from the contact C (sp. 466 and 469, Table I),

TABLE I

Contact	A	A	A	B	B	B	C	C	
Specimen	424	426	423	449	451	455	466	469	432
Modal Analysis.									
Quartz	68	39	35	82	84	25	73	10	23
K feldspar	28	9·9	—	15	12	22	24	41	36
Plagioclase	—	35	50	—	2·4	40	0·5	39	29
Hornblende	—	—	—	—	—	4·9	—	3·6	4·9
Biotite	0·4	13	13	0·4	0·1	7·8	2·0	5·5	6·3
Muscovite	3·0	0·7	0·4	2·0	0·9	—	—	—	—
Iron Ore	0·1	n.d.	0·04	0·8	0·3	0·1	0·6	0·4	0·2
Apatite	n.d.	n.d.	0·09	—	—	n.d.	—	0·09	0·01
Sphene	—	—	—	n.d.	—	0·4	n.d.	0·3	0·3
Colour Index	3·6	14	13	2·9	1·3	14	2·7	10	12
Values Recalculated to 100 per cent.									
Quartz	70	46	41	84	85	29	75	11	26
K feldspar	30	12	—	16	12	26	25	46	41
Plagioclase	—	42	59	—	2	46	0·5	43	33
Calculated Chemical Composition (Weight per cent).									
SiO_2	87	71	70	93	93	64	85	64	68
Al_2O_3	6·3	13	15	3·5	3·3	18	6·3	19	16
(Fe, Mg)O	0·5	4·5	4·6	1·1	0·05	4·6	0·9	3·7	3·0
CaO	—	1·5	2·1	—	0·1	2·9	0·3	2·3	2·1
Na_2O	—	3·3	4·5	—	0·3	4·4	0·06	3·8	3·0
K_2O	4·9	3·2	1·6	2·7	2·2	5·0	5·4	7·2	6·6
Fe_2O_3	n.d.	n.d.	n.d.	n.d.	0·6	0·02	1·6	0·8	0·4
von Wolff Values.									
L	27	44	58·5	15	14·3	87	31	84	73
M	1	16	10·5	1	0·3	12	2	7	9
Q	72	40	31	84	85·3	1	67	9	18

(n.d. = not determined.)

and contacts A and B may be combined to provide a fairly complete section across the contact. Consequently all the data may be shown on a single pair of diagrams (Text-fig. 3 (*a*) and (*b*)). From a study of these diagrams and the textures of the appropriate thin-sections it is possible to obtain an accurate impression of the progressive



TEXT-FIG. 3.—(*a*) Composite partial mode diagram of Bloody Foreland contact.
(*b*) Composite von Wolff diagram of Bloody Foreland contact.
466 quartzite 1¼ feet from granite contact C.
469 granite 4½ feet from quartzite contact C.
Other specimens as in Text-figs. 1 and 2.

mineralogical and chemical (Text-fig. 4) changes across the contact. It is significant that many analogies may be drawn between these diagrams and those, based upon actual chemical analyses, illustrating Reynolds' (1946) paper on the sequence of geochemical changes leading towards granitization.

2. The study of the Bloody Foreland contact provides results analogous to those obtained by chemical analysis in other areas, but it is difficult to assess the absolute accuracy of the present method from this example. However, a more rigorous test of accuracy has been applied in the study of the modal variation of the granite throughout Gweedore. Modes were determined of forty-five granite specimens which have been collected to be as representative of the whole area as possible. An arbitrary 1 sq. mile grid was constructed for the area, and where more than one mode was available for a square the values were averaged. The results were then plotted at the centres of the grid squares on a map and the values "contoured"; the following values were investigated in this way:—

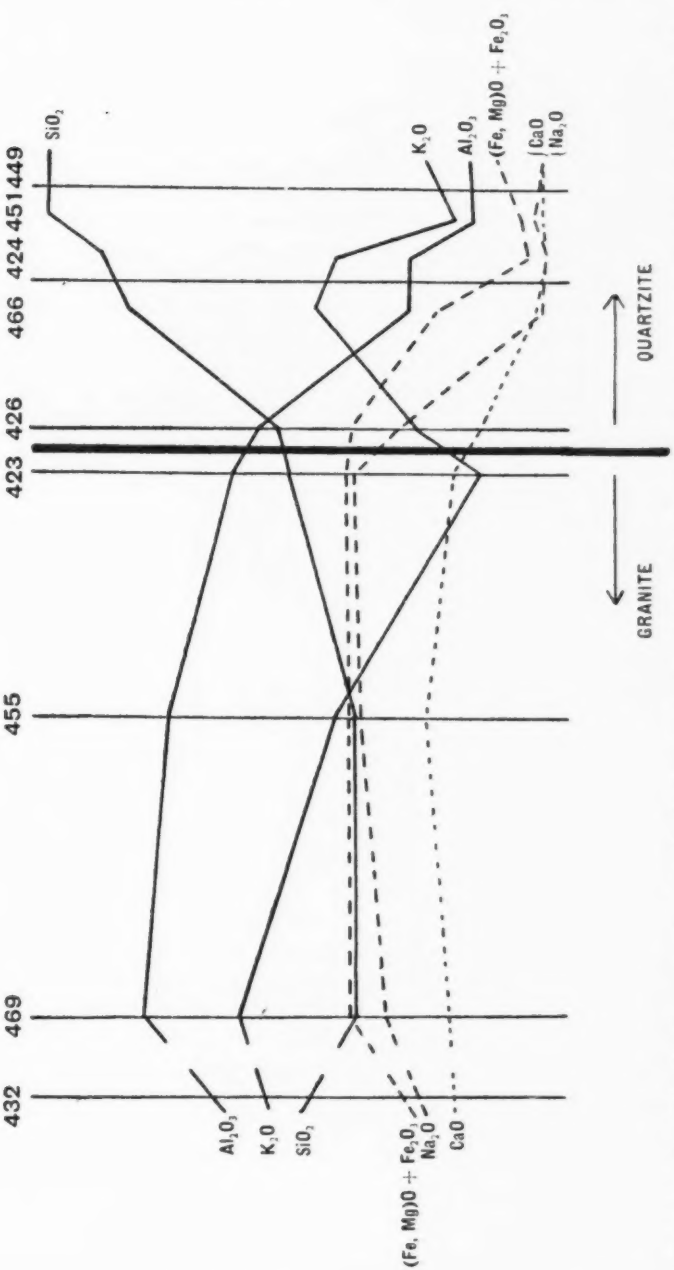
(*a*) total modal quartz;
(*b*) colour index;
(*c*) total modal feldspar;
(*d*) ratio of potassic feldspar to plagioclase.

The degree of consistency and harmony shown by these results is very remarkable, and seems to justify the claim that a fairly accurate picture of the variation within the granite-gneiss may be obtained by this simple method.

Conclusion

The main conclusion to be drawn from this work is that the accuracy of the modal analyses and derived chemical analyses is very consistent. It is hoped to publish the full petrographic data concerning the rocks discussed in the near future. It is not claimed that individual modes calculated from single slides give highly accurate results. Rather, the modes of a suite of analogous rocks give a reasonable demonstration of the variability existing in the suite as a whole. Likewise, a sufficiently accurate representation of the chemical variation may be obtained from these modes provided that the following two sources of error are borne in mind:—

(*a*) The chemical composition of the minerals involved will not be exact. Although this is a real and significant source of error in some cases, for any one series of contacts the error is probably roughly consistent, i.e., an individual mineral species will probably not vary significantly in chemical composition from one part of a granite-quartzite contact to another. This may not be a reasonable conclusion



TEXT-FIG. 4.—Calculated progressive chemical changes across the Bloody Foreland granite-quartzite contact. (Vertical scale variable.)

where complex silicates and minerals with marked rhythmic zoning are present in a rock.

(*b*) The initial mode will be inaccurate. In examining a volume of rock, or the contact of two rock bodies, a complete three-dimensional volumetric study is normally impossible, and hence it is necessary to extrapolate from data that are essentially two-dimensional. Thus, even if extremely accurate modal (or chemical) data could be deduced for individual rock samples, it would be erroneous to suggest that they constituted more than an approximation of unknown accuracy when the rock mass as a whole is considered. It is contended that by the application of an uniform technique such as outlined in this paper, a sufficiently accurate assessment of the broad variations extant in a complex may be made.

The expenses of this work have been defrayed by a grant from the Central Research Funds of the University of London, which is gratefully acknowledged.

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Critical Sections in a Carboniferous Reef Knoll

By W. W. Black

A study of the structure and petrography of the limestones exposed at Hall Hill, in Bowland, enables the geometry of part of a Carboniferous reef knoll to be defined. The main features are a central mound-like core of unbedded, poorly fossiliferous calcite mudstone and a peripheral area of bedded calcite mudstones with original, depositional dips. Thick, well-bedded coarse crinoidal limestones were subsequently bedded over the underlying knoll.

CARBONIFEROUS reef limestones in the north of England have attracted much attention during the last sixty or so years, because of their fossils, which are sometimes abundant and beautifully preserved, and because of the stratigraphical and structural problems which they present. Tiddeman (1889, 1892) originally drew attention to the ovoid or conical knoll topography usually associated with these limestones and suggested that this was the form in which the limestones accumulated. It followed, according to Tiddeman, that the dips to be seen on the flanks of the knolls were also original features of deposition and were due to the sloping surfaces on which the limestone sediment had been laid down. Both these views were challenged by J. E. Marr (1899), who preferred to postulate a tectonic origin, and there is still no general agreement on the matter. Papers by Parkinson (1935, 1944, 1950), Hudson (1932), and Bond (1950*a*, 1950*b*) are a good guide to recently expressed views.

The origin of the dips in these limestones is difficult to prove conclusively to anyone who is not sympathetic to one's own views on the subject. It is obvious that even if depositional dips are present these may have been modified by subsequent tectonic activity and it then becomes necessary to eliminate the tectonic element. This can only be done with certainty where not only the reef limestones but also the underlying non-reef limestones [1] are perfectly exposed together. The dip of the non-reef limestones may then be accepted as a record of tectonic activity and if the dip in the overlying reef limestones is significantly different from this then that difference must be a depositional feature. Since it is common experience that dips may vary significantly in exposures a few yards apart it is necessary, for complete certainty, that both facies should be seen in one exposure. Actual examples of this kind are easiest to find where the knolls are very small and Dixon (1921), Morris (1929), Shirley and Horsfield (1944), and the writer (1950) have drawn attention to some of these. In the

[1] The terms " reef limestone " and " non-reef limestone ", as used here, are defined by the full descriptions of each given later in this paper.

larger knolls, such as those of Bowland, in the Craven Lowlands, conclusive exposures of this type are rare and it is the purpose of this paper to draw attention to what is probably the best of these, hitherto unrecorded.

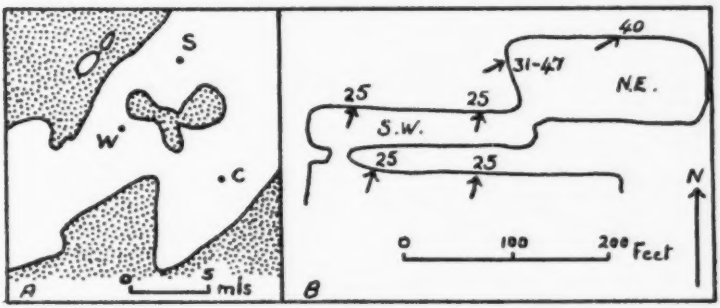
The knolls of the Clitheroe, Slaidburn, and Little Bowland areas of the Craven Lowlands have been the subject of a series of papers by Parkinson (1926, 1935, 1936, 1944, 1950) who has vigorously supported many of Tiddeman's views on their origin. In his paper on "The Geology and Topography of the Limestone Knolls in Bowland" (1935) he deals with those found in the Whitewell area, giving a general description of the stratigraphy in addition. He concludes that the knoll topography and the dips in the reef limestones are original features, a conclusion which the writer's work in the same area strongly supports.

The Whitewell area lies about 6 miles to the west-north-west of Clitheroe and consists of Carboniferous rocks ranging from the zones C to E_1. Structurally, the area lies in the zone of north-east to south-west trending Bowland folds, dips in the competent beds ranging from 10 to 45 degrees. The lowest exposed beds belong to the Clitheroe Limestone which contains an extensive development of reef limestones near its top. Knoll topography is well developed in these reef limestones in which exposures, though numerous, are usually small. There is, however, one group of exposures which penetrate into the base of a knoll and which enable definite conclusions to be reached about its geometry.

The knoll in question is known as Hall Hill and lies about half a mile east of Whitewell village. It has been extensively quarried along its south face and in these quarries (Text-fig. 1), the lowest beds of the reef limestones are clearly seen together with the underlying non-reef facies. The latter are well-bedded, dark limestones with shale wisps and partings and with some highly crinoidal beds; chert is fairly common. The macro-fauna is small apart from crinoid debris. Occasional zaphrentoid corals and productids occur, together with echinoid plates which are locally quite common. The micro-fauna, however, is abundant and consists of foraminifera, *Endothyra* and *Rhizomina* being the commonest. In thin section the limestone is seen to consist of a varying amount of crinoid debris and irregularly shaped calcite crystals together with a small amount of dark calcite mud which is often limonitic. The rock is often roughly banded with coarser and finer layers alternating. Rounded, pellet-shaped grains of calcite mud are sometimes common and may be interpreted as excretions. Chalcedonic silica occurs in small quantities often replacing part of a crinoid ossicle.

The reef limestones are in part unbedded, but in general show fairly

well developed bedding which can be clearly distinguished from the jointing. They consist of very fine-grained limestones which are often true calcite mudstones or porcellanous limestones and which weather light grey or white. Bands of recrystallized calcite often referred to as "tufa" are common and have been discussed in a previous paper (Black, 1952). In thin section the rock is seen to consist largely of semi-opaque calcite mudstone which has, in places, been recrystallized. The tendency of the rock to break along this recrystallized material often obscures the true lithology of the limestone in macro-examination. Authigenic pyrite and quartz are common, particularly the latter, which occurs in well formed bipyramidal crystals with carbonate inclusions. These crystals may make up as much as



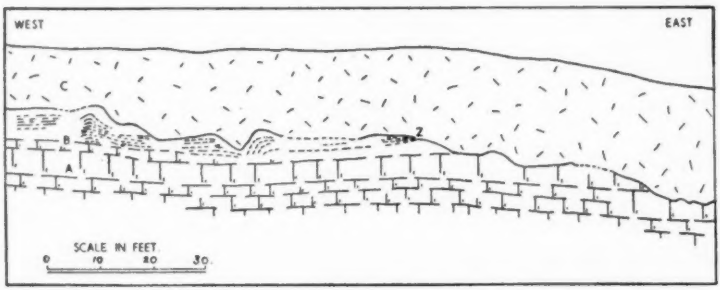
TEXT-FIG. 1.—A. Sketch-map showing position of Whitewell (W), Slaidburn (S), and Clitheroe (C). Stippled areas are Millstone Grit ; blank areas Lower Carboniferous.

B. Sketch-plan of quarries on the south side of Hall Hill.

20 per cent of the rock by weight and are sometimes arranged in definite bands which give the limestone an appearance of lamination. The micro-fauna is variable in amount and consists mainly of the uniserial bryozoan *Polypora* and occasional *Calcisphaera*. The macro-fauna is of the characteristic reef-type which has been described by many workers (e.g. Parkinson, 1926, pp. 200–210). Small-scale banded structures, possibly of algal origin, are occasionally seen, but on the whole are rare.

The relationship of reef and non-reef limestones is clearly seen in the south-western part of Hall Hill Quarry (Text-figs. 1 and 2). Here the main face runs east–west and is about 80 yards long. At the western end the lowest 18 feet are of non-reef limestone dipping 010° 25° and these beds are overlain by unbedded reef limestones. As the non-reef limestone is traced eastwards along the face the beds pass successively into porcellanous reef limestones, commencing at the top.

Thus the entire 18 feet is represented by porcellanous reef limestone at the eastern end. The actual passage is very sharp and at first sight the reef limestone appears to be markedly unconformable upon the non-reef, cutting out successive beds as the contact is traced from west to east. The appearance of unconformity is further enhanced by small drag-folds in the shaly top of the non-reef beds—folds which cannot be traced upwards into the overlying reef limestones. Where a physical break occurs between the two facies it is seen to be an irregular movement surface carrying abundant slickensides and reflecting the very different behaviour of the two facies under compressive stress. The appearance of unconformity, however, is quickly dispelled when the beds are examined in detail. The non-reef limestone is seen to change gradually as the sharp contact is approached. The shale bands and the



TEXT-FIG. 2.—Diagrammatic section in the south-west quarry, Hall Hill. A is thickly bedded non-reef limestone ; B is thinly bedded non-reef limestone with shale partings and showing drag folding ; C is unbedded reef limestone. The passage from non-reef to reef facies is well seen at Z (see Text-fig. 3).

chert nodules disappear; the rock becomes more massive and takes on a lighter colour; and elongated, irregular patches of porcellanous limestone occur near the actual contact. Nevertheless, the characteristic micro-fauna is maintained up to the contact as is shown by thin-sections cut across it. In some parts of the face the non-reef limestone passes into porcellanous reef limestone with no physical break and with considerable increase in thickness of individual beds.

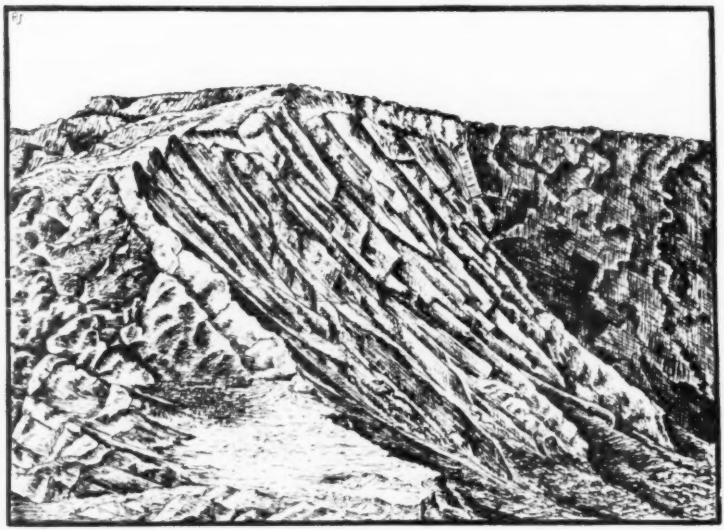
The north-eastern part of the quarry consists entirely of reef limestones, well-bedded and containing occasional highly crinoidal beds, which dip off the unbedded reef limestones at 030° 30–47°. Below the unbedded reef limestones are the non-reef limestones dipping at 010° 25°. This disparity of dip is the key to the entire structure. The dip of the non-reef limestones is the same as that found elsewhere in the vicinity and is clearly tectonic in origin. The dip of the bedded reef limestones is compounded of this tectonic element plus a non-

tectonic element which is found by subtraction to vary between 13 and 28° to the north-east. Before this non-tectonic element can finally be accepted as an original dip of deposition it is necessary to consider possible alternative explanations. It is possible, for example, that two sets of beds showing different dips might be separated by a thrust fault. In the present case support for such a view could be obtained from the presence of drag folds, calcite veining and extensive slickensiding at some of the contacts between reef and non-reef facies. It can, however, be stated quite definitely that, since continuous passage (i.e. with no fracture of any kind) can be seen at a number of points, the amount of movement which has occurred at the contact is small and developed on a number of discontinuous surfaces. All the features of movement which are present are adequately explained by the differing competency of reef and non-reef facies.

Differential compaction offers another attractive possibility which has been discussed in relation to reef structures by a number of workers (see, for example, O. T. Jones (1944, p. 150), Parkinson (1950, p. 344)). This mechanism involves the presence of a non-compactable core or central area of the reef and a highly compactable periphery. The latter is due to interdigitation or admixture of muds with a high initial water content, the subsequent loss of water as the sediments are buried causing a diminution in thickness at the periphery but not at the core. This results in outwardly directed dips at the margins of the reef. At Hall Hill, however, the absence of interdigitation or admixture of argillaceous sediments in the beds under consideration rules out such an explanation. The limestones are remarkably pure, showing 98 per cent $CaCO_3$ and only 0·16 per cent Al_2O_3, and are identical in lithology with those in other parts of the reef. The presence of thin shelled, hollow, and perfectly preserved fossils argues against any appreciable compaction of the original calcareous sediment.

There is, however, evidence which strongly supports the view that the non-tectonic element of dip is depositional. In the north-eastern part of the quarry certain beds show a marked increase in thickness when followed up the dip from the floor of the quarry to its top—i.e. in a south-westerly direction (see Text-fig. 3). The increase is from 2 ft. 6 in. to 5 ft. 6 in. in 35 ft. and takes place not only by the thickening of individual beds but also by the incoming of new wedge-shaped beds. This is incontrovertible evidence of lenticular or knoll-like deposition. As a result of this, the upper surface of this group of beds has a slightly higher dip than the base, but, in addition, there is an overall increase in dip from 31° at the quarry floor to 47° at the quarry top. This represents a change from 13° to 28° in the depositional dip as the beds are traced towards the centre of the knoll, the thinner beds at the margin having been deposited at a lower angle than the thicker beds nearer

the centre. Also, at the maximum observable thickness of this group of beds the bedding becomes obscure. Since thickening implies more rapid and therefore possibly more continuous sedimentation, this is not unexpected. Conversely, the slower rate of sedimentation towards the margin of the knoll is consistent with occasional pauses in deposition and the consequent development of bedding planes. It follows, therefore, that these beds were deposited on a slope of between 13° and 28° facing north-east and that this slope formed part of a limestone



TEXT-FIG. 3.—View of Hall Hill Quarries from the south-east. A part of the south-west quarry is seen on the left showing, at the base, non-reef limestones dipping at 010 25°. These are overlain by unbedded reef limestones and also pass laterally into reef limestones in the foreground. The bedded reef limestones form the centre of the sketch and dip to the north-east at angles varying from 31° to 47°. Note the increase in thickness and in angle of dip as the beds are traced from the floor to the top of the quarry, a distance of about 35 feet. (Drawn by Mr. R. Johnson from a photograph.)

knoll the thickest part of which was situated to the south-west of the quarry. Since the south and west margins of Hall Hill are recent erosion features it is clear that the knoll must originally have extended for a considerable distance in this direction.

The facts outlined above allow us to reconstruct a segment of a growing reef-knoll during a brief interval of time. They show that a calcite mud, deposited on a slope varying from 13° to 28°, was not only able to resist the force of gravity, which might have been expected

to cause slumping and sliding, but that the angle of slope was slightly increased as a result of this deposition. Clearly, some kind of reef-building organism would offer the simplest mechanism to bring this about, but there is no indication that such was in fact present. If it existed it has left no trace observable by the writer either in the exposures or in thin sections. Neither does the reef limestone consist of a mound of shell debris, as originally suggested by Tiddeman, since large masses of the limestone are virtually unfossiliferous—a fact which has been noted elsewhere by other workers and has been discussed recently by Parkinson (1950, p. 346). Measurements made on nineteen thin-sections from Hall Hill quarry show that the reef limestones contain an average of 12 per cent of shelly material whereas the non-reef limestones contain an average of 34 per cent. The reef limestone is less fossiliferous (in the strict sense of the word) than its non-reef equivalent and consists dominantly of calcite mudstone, often recrystallized to a varying degree. The higher beds of the Hall Hill knoll (outside the quarry) are markedly crinoidal, as is often the case with the higher beds in other knolls. As recently noted by Parkinson (1950, p. 347), however, crinoidal limestones are not an essential component of these knolls and their presence can have little bearing on the problem of knoll formation: this problem revolves around the origin of the calcite mudstone.

When comparing the reef limestone with its non-reef equivalents the most striking point which emerges is that the latter is markedly clastic, the former dominantly non-clastic. Even where the reef limestones are rich in crinoid debris this is unabraded and shows no sign of transport. It has in fact accumulated more or less in position of growth and has been bedded over the underlying knoll structure. Such a deposit would have the physical properties of a coarse sand and would form stable deposits on slopes of the order of 30° in the absence of strong currents able to transport the crinoid debris. It is, however, more difficult to account for the accumulation of calcite muds on slopes up to 28° in the apparent absence of a reef-building organism. Absence of strong currents again seems to be necessary and cementation synchronous with deposition is a possible mechanism. The origin of the calcite mud remains an open question, but it seems to the writer that, unless more evidence emerges, it can only be regarded as a direct precipitate, perhaps involving bacterial activity.

These theoretical considerations are of great interest but it is the concrete geometrical evidence to be found in the Hall Hill exposures which should be emphasized. This evidence demonstrates, with as much finality as seems possible, the essential truth of Tiddeman's views on the original nature of the knoll form and of the presence of depositional dips in the reef limestones.

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Petrology of Chloritoid and Staurolite Rocks North of Stonehaven, Kincardineshire

By Douglas H. Williamson

Abstract

In the pelitic rocks of this classical section it appears that, chemically, a Fe_2O_3: Al_2O_3 ratio of about 0·40 is critical for the formation of chloritoid or staurolite. With this ratio both minerals are possible, but if the value exceeds a figure somewhere between 0·47 and 0·55 staurolite cannot form, a reddish-brown biotite appearing instead. Physically, a certain minimum temperature is necessary before staurolite can appear and below this chloritoid is the equilibrium phase.

METAMORPHIC rocks carrying staurolite and chloritoid have been described by many writers (Barrow, 1893; Barth, 1936; Harker, 1932; Read, 1934; Tilley, 1925), but the precise relations between these two minerals are still somewhat obscure. One of the earliest descriptions of their occurrences is given in Barrow's pioneer work in 1893 on the Dalradian rocks between Stonehaven and Aberdeen. These rocks were mapped again by Campbell for the revision of the 1-inch Geological Map, Sheet 67. In the course of further detailed mapping during the last four years by this Department, the problem has been re-examined in a study of the structure and metamorphism of the region. As is well known, in Kincardineshire the pelitic members of the Dalradian north of the Highland Border Fault provide a particularly good example of metamorphic zoning in that, proceeding northwards, there is a progressive increase in the grade of the metamorphic mineral assemblages. Thus the argillaceous member, which at the Fault is represented by a low-grade phyllite, passes first through a biotite-schist, then a chloritoid-garnet-schist, until 2 miles beyond the Fault, it has become a thoroughly crystalline staurolite-schist. The present paper is concerned principally with these three groups of rocks; viz. (1) those characterized by chloritoid, (2) those containing staurolite only, and (3) an intermediate group containing both minerals.

Chloritoid Rocks

The earliest appearance of chloritoid occurs in the vicinity of the three post-metamorphic porphyry intrusions in Perthumie Bay; that is, about half a mile north of the Highland Border Fault. Here they are scattered about in the rock as irregular poeciloblastic grains, often pseudomorphed by kaolinitic material, which northwards rapidly develop into larger idioblastic crystals. This first appearance of the mineral is in the biotite zone, but it is noteworthy that as the chloritoid develops, the biotite, which has been of a green variety, and therefore

presumably poor in FeO, rapidly diminishes in amount. By the time the chloritoid reaches its best development, the biotite has completely given place to garnet, the schists now being definitely in the garnet zone. It would thus appear that chloritoid and biotite do not form a stable assemblage.

The first occurrence of fresh and well-formed crystals lies about 75 yards north of the three porphyry intrusions in Perthumie Bay already mentioned. Here its most striking development is in a rock which has a knobbly weathered surface resembling some of the coarse Dalradian grits. These knobs, which measure nearly a centimetre in diameter, are large porphyroblastic chloritoids set in a fine-grained micaceous matrix. When fresh, the crystals, which have a slaty blue colour and waxy lustre, exhibit a good basal cleavage, often curved. The habit is typically short, stunted prisms, which give near circular basal sections whose orientation is completely random, bearing no relation to the schistosity of the host rock. The mineral is sometimes accompanied by small garnets. In thin section the chloritoid crystals, which contain abundant inclusions, mostly of tiny quartz grains, are markedly pleochroic. The optical properties are:—

α	grey green	1·721	
β	slate blue	1·725	$(\beta > \alpha > \gamma)$
γ	yellowish—colourless	1·728	
	$\gamma - \alpha =$	0·007	
+	2V ca 50° $\gamma \wedge c =$	20°	

Lamellar twinning is common. It is noteworthy that these large chloritoids are often surrounded by a zone of transparent material consisting of an aggregate of minute scales of a kaolinitic mineral of very low double refraction and small quartz grains, the latter having obviously once been inclusions. At times, destruction of the chloritoid has been practically complete, so that nothing but a pseudomorph of kaolinitic material remains, with here and there a relic of the original mineral (Text-fig. 1).

Rastall (1910) has suggested that, in the case of the Skiddaw granite, chloritoid belongs to an earlier period of dynamic metamorphism and is subsequently destroyed by the contact action of the later granite. This, however, is not substantiated by the present evidence, which shows clearly that the intrusions have no effect whatsoever on the occurrence of the chloritoid, the best examples lying only 50 yards north of these bodies.

Transition from Chloritoid to Staurolite-bearing Rocks

About 800 yards north of the three porphyry intrusions in Perthumie Bay the argillaceous member shows small white spots many of which

are small narrow prisms. In thin section these are seen to be staurolite pseudomorphs some still containing a few relics of the original mineral: the characteristic form and twinning are much in evidence. The replacing material is a micaceous aggregate of very low double refraction similar in many respects to that which elsewhere replaces the chloritoid, and is in marked contrast to the replacement material of staurolite further north, which consists of a micaceous aggregate of high double refraction.

TEXT-FIG. 1.—Porphyroblasts of chloritoid with small garnets set in a very fine-grained matrix of white mica, chlorite, and quartz. Three of the chloritoid crystals show lamellar twinning and the others exhibit the usual rim of kaolinitic material and minute quartz grains.

Accompanying the staurolite, but no longer in idioblastic crystals, is chloritoid. It was hoped that the conversion of the latter to the former, or at least some textural clue to the genetic relations, might have been observed, but nowhere can staurolite be seen arising from the ruins of chloritoid as mentioned by Tilley (1925) for rocks of the Piz Scopi area. All that can be seen is that numerous allotriomorphic grains of chloritoid, either singly or gathered into patches and knots, are scattered throughout the highly micaceous matrix, the chloritoid never occurring inside the large staurolite pseudomorphs, nor forming a mantle to that mineral. It is noteworthy that biotite, now a reddish brown variety, rapidly becomes a stable phase once staurolite has become established as the principal metamorphic mineral. The appearance of this biotite, which is presumably rich in iron, would be

very much in accord with Tilley's equation for the formation of staurolite:—

Chloritoid + muscovite + quartz → staurolite + biotite + water

$$10\,H_2FeAl_2SiO_7 + H_2KAl_3Si_3O_{12} + SiO_2 \rightarrow 4HFeAl_5Si_2O_{13} + \left\{\begin{matrix} H_2KAl_3Si_3O_{12} \\ 3Fe_2SiO_4 \end{matrix}\right\} + 8H_2O$$

The main drawback to this synthesis is the complete lack of evidence in this area for staurolite actually arising from the ruins of chloritoid. Despite this development of the staurolite the mineral never occurs intergrown with the chloritoid.

For about 800 yards north of this first exposure of staurolite, a few rocks are encountered which still carry chloritoid to the exclusion of the higher grade mineral, although the majority contain both minerals with the staurolite becoming more and more predominant. It is noteworthy that in those rocks rich in staurolite the small amount of chloritoid still remaining shows a more intense pleochroism from a yellowish colour to a bright blue.

The appearance of staurolite in the schists coincides with the development of fresh albitic plagioclase. The feldspar is allotriomorphic, contains many inclusions, and forms medium-sized crystals. Of interest is the fact that these feldspars occasionally contain inclusions of chloritoid, the feldspar sometimes developing around the alteration products of chloritoid, so that there is a core of a kaolinitic mineral with fine quartz and haematite. Such cores are to all intents and purposes identical with the chloritoid pseudomorphs occurring further south in the garnet zone.

Staurolite Rocks

About one mile north of the prophyry intrusions, the stable iron-alumino-silicate is staurolite to the complete exclusion of chloritoid. It is always porphyroblastic but, as with chloritoid, retrograde effects are common, the alteration products consisting of an extremely fine micaceous aggregate of high double refraction—the shimmer aggregates of Barrow. Fresh crystals are not common so that the shimmer aggregate pseudomorphs are often all that remain to prove the original existence of staurolite. When fresh, the crystals have numerous fractures and a prominent (010) cleavage: the optical properties are:—

$\alpha = 1{\cdot}738$ colourless
$\beta = 1{\cdot}742$ pale yellow
$\gamma = 1{\cdot}749$ golden yellow
$\gamma - \alpha = 0{\cdot}011$; $+2V$ ca 85°

General Considerations

In all the rocks examined there is no clear textural evidence of the manner in which chloritoid has arisen. A typical environment in

which it appears to thrive consists of chlorite, muscovite, and quartz. In the early stages it is often poeciloblastic, thus indicating crystallization and growth at isolated centres which later coalesce to form a porphyroblast. In the present case, chloritoid has certainly not arisen by a reaction involving andalusite, as envisaged by Tilley (1925) for the Frenchman schists of South Australia:—

$$Al_2SiO_5 + Fe(OH)_2 \rightarrow H_2FeAlSiO_7$$

Tilley's second equation involving muscovite, would appear to be applicable since there is an abundance of white mica in the schists:—

$$3Fe(OH)_2 + 2H_2KAl_3Si_3O_{12} \rightarrow 3HFeAl_2SiO_7 + 2KOH + 3SiO_2 + H_2O.$$

The manner in which the ferrous iron exists is not clear, but chemical analysis of the schist shows a high content of FeO. If, as the equation indicates, muscovite is to be utilized for the production of chloritoid, one might expect to find it diminishing as the latter increases. This is not the case, however, as muscovite is the one mineral whose growth is being continually promoted and which is in complete equilibrium with the conditions prevailing during all phases of the metamorphism.

Harker gives an equation involving kaolin and magnetite:—

$$3H_4Al_2Si_2O_9 + Fe_3O_4 \rightarrow 3H_2FeAl_2SiO_7 + 3SiO_2 + 3H_2O + O$$

It has earlier been noted in this communication that the chloritoid often exists as pseudomorphs, the replacement material being a kaolinitic mineral along with iron oxide; that is, retrograde changes can certainly cause the equation to go from right to left. If this is a reversible reaction, and a stable equilibrium assemblage under retrograde conditions consists of a kaolinitic mineral and iron oxide, then during progressive metamorphism it would seem reasonable that chloritoid could arise by a reaction involving kaolin and iron oxide. Chemical analysis shows that the original argillaceous sediment was rich in alumina and iron and, in terms of minerals, the alumina could exist as a hydrated alumino-silicate of a platy nature such as is commonly found in argillaceous sediments. It is therefore suggested that during the metamorphism the production of chloritoid by reaction between the platy hydrated alumino-silicate and iron would involve a minimum rearrangement in the mineral structure, since both are platy minerals. The process would essentially be one of dehydration with iron replacing the water.

An analysis of a chloritoid schist is given in Table I together with the compositions of other chloritoid schists for comparison.

From this it will be seen that chloritoid schists in the present area are richer in alumina and iron than those given for comparison, with the exceptions of the chloritoid schists from Unst and New York

State. All are relatively poor in lime and magnesia, whilst potash is fairly high indicating an abundance of white mica. It will be seen that silica percentages may vary widely, a point at once confirmed by the fact that chloritoid, and also staurolite, sometimes occur in gritty rocks, consisting almost essentially of quartz pebbles.

As has already been stated there is no textural evidence in the present area that staurolite develops at the expense of chloritoid, but there is a transitional zone where the two minerals co-exist. In this transitional zone, as the metamorphism increases northwards, chloritoid rapidly diminishes in amount until ultimately only staurolite is present. Thus, though the two minerals show an inverse relation to one another, the reactions involving the individual grains appear to

TABLE I

	(1)	(2)	(3)	(4)	(5)	(6)	(7)
SiO_2 .	55·26	73·54	52·86	49·53	76·90	78·88	74·48
Al_2O_3 .	20·46	14·40	25·29	26·06	15·02	10·75	15·96
TiO_2 .	0·91		1·20	1·62			
MnO .	0·15	2·37	0·13	0·17			
P_2O_5 .	0·44		0·17	0·12			
Fe_2O_3 .	1·89	4·59	1·22	3·00	0·50	1·05	0·87
FeO .	7·12		7·83	7·78	4·72	2·74	2·02
MgO .	2·81	0·50	1·67	2·59	0·58	1·67	0·35
CaO .	0·73	0·02	0·25	1·06	0·50	0·10	0·28
Na_2O .	2·44	0·44	2·25	1·03	0·11	1·26	0·72
K_2O .	3·17	2·11	3·62	3·78	0·01	1·85	3·88
H_2O+ .	3·74	1·96	3·94	3·09	1·30	1·49	2·02
H_2O- .	0·77	0·24	0·10	0·09	0·20	1·20	0·13
Totals .	99·89	100·17	100·53	99·92	99·84	100·99	100·71

(1) Chloritoid-schist north end of Perthumie Bay. Ground mass of quartz, muscovite, chlorite, and iron ores. Analyst: D. H. Williamson.
(2) Quartz-sericite-chloritoid-schist. Gotthard massiv. Niggli.
(3) Chloritoid-schist. S.E. of Crouses Store, Clove quadrangle, New York State. Analysts: Ellestad and Barth.
(4) Chloritoid-schist, Snarravoe, Unst. Analyst: B. E. Dixon.
(5) Chloritoid-schist, Potaro River, Guiana. Harrison, 1908.
(6) Chloritoid-schist, Alpi Apuane, Manasse. Niggli.
(7) Chloritoid-schists. The Frenchman, Eyre Peninsula, S. Australia. Tilley, 1925.

proceed quite independently. In its earliest formation the staurolite is always porphyroblastic and set in a matrix which in every way is similar to that from which chloritoid develops. Various reactions have been given for its formation. Harker (1932) gives a reaction between chloritoid and kyanite: Pelikaan (1913) mentions a straight conversion of chloritoid to staurolite, the liberated iron oxide converting muscovite to biotite; Tilley (1925) advocates a reaction using chloritoid and muscovite. Unlike the Stonehaven area, all the above cases show mineralogical evidence for the utilization of chloritoid in the production

of staurolite. It would therefore appear that in the present case, staurolite must develop from the essential constituents of the ground mass itself of the argillaceous sediments.

The alumina necessary for the foregoing reactions must exist in the ground mass material. Chemical analyses given in Table 2 of two staurolite rocks from this area show that there is a large increase in alumina, as much as 8 per cent, over a chloritoid rock from the same district. Thus one important factor in the production of staurolite is richness in alumina, although there must also be an adequate source of iron. Chemical analyses of staurolite and chloritoid show a ratio of Fe_2O_3: Al_2O_3 of approximately 3: 10 for the former and 5: 8 for the latter mineral.

TABLE 2

	(8)	(9)	(10)	(11)
SiO_2	49·16	56·11	55·79	44·13
Al_2O_3	28·21	18·64	25·17	26·13
TiO_2	1·01	1·25	1·51	1·11
MnO	0·19	0·10	0·10	0·18
P_2O_5	0·47	0·49	0·04	0·52
Fe_2O_3	4·63	6·60	7·07	3·63
FeO	6·21	3·93	2·53	8·43
MgO	1·84	2·13	1·93	3·07
CaO	0·88	0·37	0·68	0·95
Na_2O	3·00	2·64	1·12	3·11
K_2O	1·83	4·18	3·00	4·79
H_2O+	2·76	3·45	1·29	3·59
H_2O-	0·28	0·10	0·10	0·77
Totals	100·47	99·99	100·33	100·41

(8) Staurolite-schist, Limpet Burn, Kincardineshire. Analyst: D. H. Williamson.

(9) Biotite-schist, Limpet Burn, Kincardineshire. Analyst: D. H. Williamson.

(10) Staurolite-kyanite-garnet gneiss, Libbers Hill, Unst. Analyst: B. E. Dixon.

(11) Staurolite-schist, Muchalls, Kincardineshire. Analyst: D. H. Williamson.

The bulk composition of the rock thus appears to decide whether staurolite can develop. Analysis No. 9 supports this point, in that the rock, although occurring well within the staurolite zone, is a biotite schist with no staurolite, and is relatively low in alumina.

From Table 3 prepared from analyses of argillaceous rocks in the area under review, it can be seen that the ratios Fe_2O_3: SiO_2 and Al_2O_3: SiO_2 are of no great consequence in determining the type of iron alumino-silicate that can appear.

TABLE 3

	(*a*)	(*b*)	(*c*)	(*d*)	(*e*)	(*f*)
$Al_2O_3 : SiO_2$	0·32	0·43	0·37	0·57	0·59	0·33
$Fe_2O_3 : SiO_2$	0·13	0·16	0·16	0·22	0·27	0·19
$Fe_2O_3 : Al_2O_3$	0·41	0·38	0·44	0·38	0·46	0·56

Total iron has been calculated as Fe_2O_3 due to refractory nature of staurolite.

(*a*) Typical phyllite Skatie Shore.
(*b*) Biotite-schist, south end of Perthumie Bay.
(*c*) Chloritoid-garnet-schist, north end of Perthumie Bay.
(*d*) Staurolite-schist. Limpet Burn.
(*e*) Staurolite-schist, Muchalls.
(*f*) Biotite-schist from within staurolite zone. Limpet Burn.

So far as the chemistry alone is concerned, in the present series of rocks staurolite seems to demand a ratio of $Fe_2O_3 : Al_2O_3$ of about 0·40. Within the same range chloritoid can form, as shown by a comparison of the $Fe_2O_3 : Al_2O_3$ ratios of (*c*), a chloritoid rock, and (*d*) and (*e*) staurolite rocks. Given the necessary physical conditions then, the pelitic rocks (*a*)–(*e*) are all capable of carrying chloritoid or staurolite. But it would appear that if the ratio $Fe_2O_3 : Al_2O_3$ exceeds a figure somewhere between 0·47 and 0·55, staurolite will not form, biotite appearing instead, as shown by (*f*) which is a biotite-schist from well within the staurolite zone. However, it must be remembered that when staurolite does appear, it is accompanied by the reddish brown variety of biotite, into which the excess iron not required for the formation of staurolite presumably enters: with chloritoid rocks, on the other hand, there is generally no biotite, so that nearly all the iron is taken up by the more ferriferous chloritoid. It would therefore appear that so far as chemical composition is concerned, either chloritoid or staurolite can form in rocks having the requisite $Fe_2O_3 : Al_2O_3$ ratio (about 0·40 for the present area), but in those rocks having a high total percentage of alumina in addition, staurolite will develop in preference to chloritoid. From the physical point of view, the order in which the minerals appear in the metamorphic zones indicates that a certain minimum temperature is required before staurolite appears. Below this critical temperature chloritoid is the equilibrium phase.

It should be pointed out, however, that elsewhere chloritoid can develop with an $Fe_2O_3 : Al_2O_3$ ratio as low as 0·18, as in a chloritoid-sericite-schist from the Alps (Niggli, 1912), and staurolite also is found with an $Fe_2O_3 : Al_2O_3$ ratio as low as 0·24 in a schist from Portsoy (Geological Sheet No. 96). The $Fe_2O_3 : Al_2O_3$ ratio can thus vary between wide limits. It therefore seems that, although the production of either mineral is to some extent influenced by the bulk composition of the rock, in that its production requires an environment rich in

iron and alumina, other factors must exercise some control, such as the local stress-temperature distribution.

I desire to record my thanks to Professor T. C. Phemister for his interest and guidance during the preparation of this paper. The work was carried out in the Department of Geology, University of Aberdeen, and is part of a larger study by the Department of the coastal section north of the Highland Boundary Fault.

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The Development of *Monograptus dubius* and *Monograptus chimaera*

By Margaret Walker

Abstract

The growth of the rhabdosome from the initial prosicula is described for two species of *Monograptus*. Sections of the metasicula have shown the nature of the "Stillstandsgürtel" first described by Kraft, and in both species evidence about the structure of the interthecal septum is given. It appears that there are two forms of *Monograptus dubius* which have the same external appearance but differ in the details of early development.

Introduction

THE material described in this paper has been obtained from a fragment of limestone found in the Pleistocene deposits of the North German Plain; the fauna shows that the rock is of Lower Ludlow (i.e. Upper Silurian) age.

The graptolites are preserved in chitin[1] and consist largely of individuals of *Monograptus dubius* (Suess) and *Monograptus chimaera* (Barrande), at all stages of growth. In the prosicula and metasicula stages it has not been possible to distinguish these forms; therefore a general description applicable to both will be given. From the initial bud onwards the two species can be readily separated.

The rest of the fauna consists of the following species:—

Monograptus scanicus, Tullberg.
M. nilssoni (Barrande).
M. bohemicus (Barrande), SM. A24375.
Retiolites mancki, Münch, SM. A24373.

The originals of the figured specimens are in the Sedgwick Museum and bear the catalogue numbers of the Sedgwick Museum Collection.

Development of the Early Growth Stages

Prosicula.

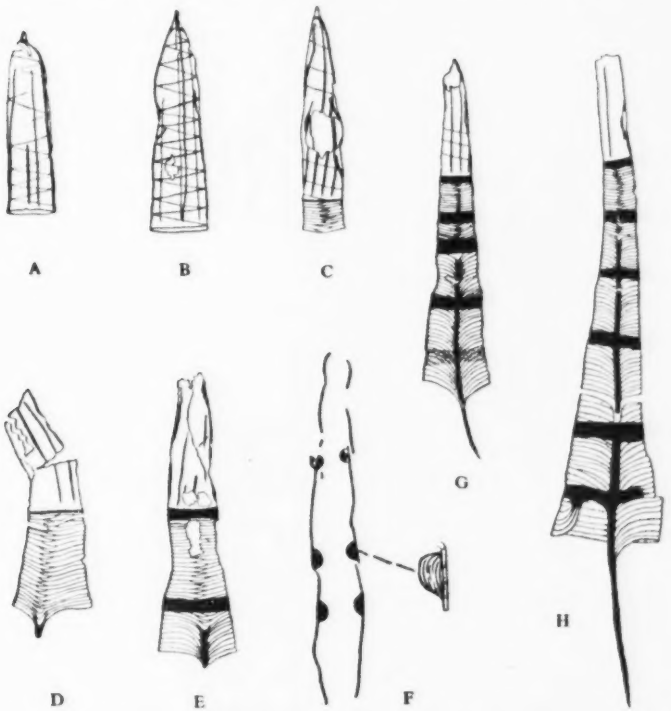
This first stage is about 0·5 mm. in length and has the usual bottle-shaped appearance, but with a very short neck part which forms a small point at the apex. The spiral line can often be seen, but is usually faint and sometimes not discernible at all; the bands are fairly evenly spaced along the prosicula (Text-fig. 1A). The earliest individuals seen show three longitudinal rods; probably these were formed subsequently although no stage without rods was found. Later prosiculae

[1] The term "chitin" is used for convenience only and does not imply exact knowledge of the substance forming the graptolite rhabdosome.

have up to about six rods which extend almost, but not quite, to the apertural margin. Three or four of these reach the nema, which is still a very short structure (Text-fig. 1B). At first the rods are well-marked; after the formation of the metasicula and early thecae, the whole of the prosicula wall is thickened and the rods become less distinct.

Metasicula.

Growth of this part begins with the deposition of half rings of chitin



TEXT-FIG. 1.—Early growth stages of *Monograptus*. A, B, prosicula; C–E, metasicula. × 50 approx. F, longitudinal section of metasicula, showing thickening rings, × 80 approx., and detail of one thickening further enlarged. G, H, metasicula. × 35 approx. SM. A24344 to 24351.

on the apertural margin of the prosicula, alternately on one side and the other. The strips overlap where they meet, giving the two zig-zag sutures which are characteristic of this and all successive stages in the rhabdosome (Text-fig. 1C). For about twenty growth-bands the strips are of constant width so that the margin of the metasicula remains nearly straight; after this the bands widen along one of the

sutures, thus forming a lobe on the margin which is the beginning of the virgella (Text-fig. 1D). Soon this structure becomes so thickened that the growth-lines on it can no longer be seen and it projects from the edge of the metasicula as a marked spine (Text-figs. 1E and 1G). With continued growth successive rings become slightly wider and increase in diameter so that the sicula has the shape of a long cone (Text-fig. 1G).

In the meantime a number of secondary thickening bands appears on the wall of the metasicula. Usually these are parallel to the growth-lines and about three or four growth-bands wide. The first one is at the edge of the prosicula and on the early part of the metasicula; the second occurs above the point where the virgella begins, perhaps about 15 to 20 growth-lines from the prosicula, and later ones are fairly evenly spaced along the length of the metasicula (Text-fig. 1G). The usual number is five (Text-fig. 1H), although specimens with six (Text-fig. 4C), four (Text-fig. 4G), or three (Text-fig. 5A) are known, and they may be oblique to a varying degree (Text-fig. 4A, F), or irregularly spaced. These are the "Stillstandsgürtel" of Kraft (1926, p. 234), but they do not mark periods of less rapid growth in the metasicula, as he thought. They are secondary: a metasicula with such a thickening at the edge is never found; the first band develops above the beginning of the virgella but is seen only in specimens where the virgella has already formed; and in some cases the thickening can be seen in its inception as a dark brown, but not completely black, band with growth-lines at the usual spacing within it (Text-fig. 1G, and oblique apertural band of 4A).

Longitudinal sections of the metasicula show clearly that this thickening material was deposited on the inside of the structure forming a ring about 10 to 15μ thick which projects into the interior of the sicula (Text-fig. 1F). At high magnifications the original material of the metasicula wall, with its growth-bands, can be seen on the outer margin (Text-fig. 1F, inset). The thickenings themselves sometimes show no internal structure, but in some there are traces of a lamellar arrangement of material composing the ring. The one at the prosicula margin is in some cases not uniformly dark, but shows a T-shaped cross-section as in Text-fig. 1F.

The foramen of the initial bud begins as a notch in the margin of the metasicula, next to the virgella. The growth-lines curve round the edge of the notch and, especially in *M. dubius*, end in a thickening of the metasicula wall which is localized on either side of the virgella. After about seven or eight bands have been laid down in this way, the foramen is completed by bands which grow down to meet the virgella again. The first of these only reaches to the margin of the foramen and has an expanded triangular end; the second touches the

virgella, curving upwards. Both these help to give the foramen a rounded margin (Text-fig. 2A, B).

Usually six or seven bands complete the metasicula and are so formed as to produce a slight projection on its margin; there are no apertural spines (Text-fig. 2C).

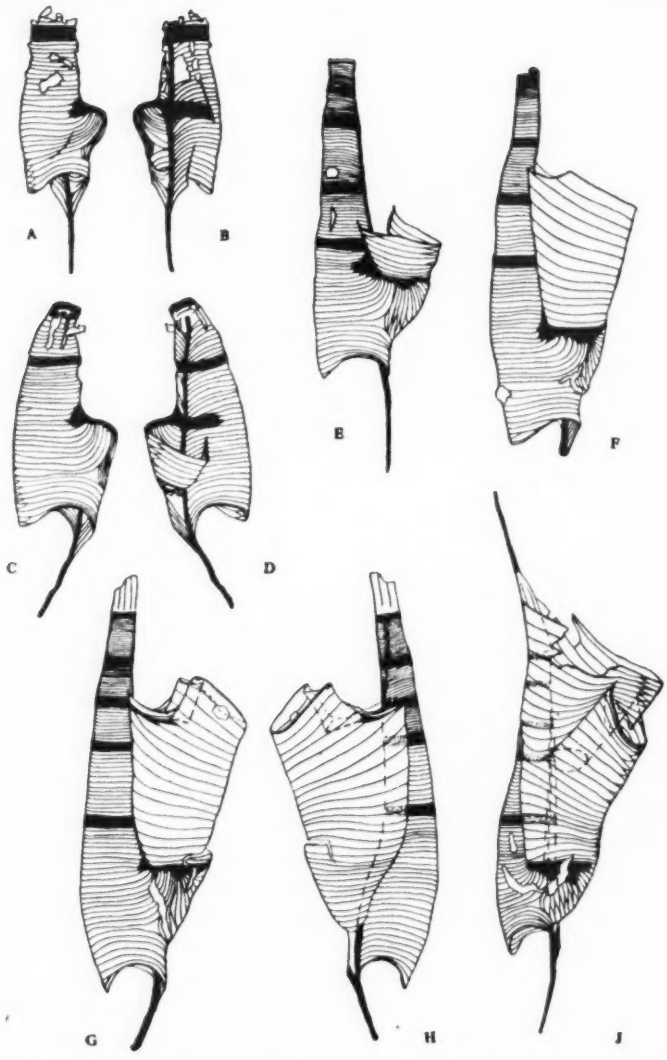
The complete metasicula is 1·5 to 1·6 mm. long (Text-figs. 4F, 2G). This is longer than the metasiculae of some diplograptids which seem to range in size from about 1·0 mm. in *Diplograptus maxwelli* (Walker, 1953 *a* and *Climacograptus inuiti* (Cox, 1933, p. 8) to 1·3 mm. in *Diplograptus leptotheca* (Bulman, 1944–7, p. 45). Since the length of the prosicula is about the same, 0·45–0·5 mm., in all these species the metasicula described here forms about three-quarters of the length of the whole sicula—a greater proportion than in the diplograptid, mentioned. This difference from the diplograptids is not constant, however, for in *M. scanicus* (Walker, 1953 *b*), the metasicula occupies only two-thirds of the total length of the sicula.

Development of *Monograptus dubius*

Initial bud and first theca.

From the metasicula growth continues with narrow strips of chitin laid along the edge of the foramen opposite the virgella, so that the bud grows laterally. Usually the first two strips are short and occupy only a part of the foramen margin, thus producing an unconformity between the metasicula and the initial bud. Later bands gradually extend along the whole length of the margin from the thickening at the top to the virgella below, and they consist of two strips joining in the middle with a zig-zag suture (Text-fig. 2A, B, C, D). The thickening mentioned above extends laterally as a prominent bar on the reverse side, with the strips from it to the suture becoming shorter as the bud grows round and upwards (Text-fig. 2C, D). When the rest of the bud has grown to the level of this thickening, the next band extends along its top edge and is attached to the metasicula, thus making what is apparently an unconformity in the middle of the growth of a theca (Text-fig. 2E). On the obverse side, however, there is no trace of the thickening; here the chitin bands join the wall of the metasicula and grow conformably for the whole of the length of the theca (Text-fig. 2H).

The first theca continues to grow upwards, the bands increasing in length so that the theca widens. They are attached to the metasicula on both sides and gradually enclose more and more of it. When the theca extends about half-way up the sicula, two bands are laid down which are free at their dorsal edges, so that they grow away from the metasicula (Text-fig. 2F). Then the chitin strips join up with each other



TEXT-FIG. 2.—Growth-stages of *Monograptus dubius*. A–D, initial bud; E–H, first theca; J, first and second thecae. All reverse side except B, D, H. A–H × 30 approx.; J × 24 approx. SM. A24352 to 24357.

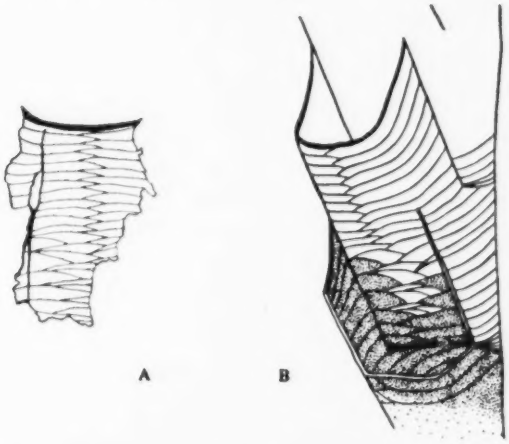
dorsally as well as ventrally, forming a complete tube. The dorsal part of this tube is the interthecal septum of the first theca and it contains the usual zig-zag suture (Text-fig. 2G, H).

About seven or eight more bands complete the first theca; it has

a straight margin which is more or less parallel to the growth-lines and is thickened by the deposition of extra chitin (Text-fig. 2J).

The second theca.

This begins with two series of chitin strips, one on each side of the rhabdosome, filling in the space between the sicula and the first theca where they have grown apart. The first few strips join the thickened upper edge of the "free" chitin bands of the first theca, so that there is an unconformity between the two thecae, and later ones



TEXT-FIG. 3.—Interthecal septum of *Monograptus dubius*. A, drawing of the ventral wall of a theca and its downward continuation as the interthecal septum of the preceding theca. A small part at the base of the septum is broken off. × 20 approx. SM. A24358. B, diagram illustrating the relation of the chitin strips in an interthecal septum to the earlier and later thecae. The ventral and left-side walls of the earlier theca are cut away, the cut edge being shown by a double line, and the parts belonging to the earlier theca are stippled.

extend from the metasicula to the dorsal part of the tube of the first theca (Text-fig. 2G, H). Above the aperture of the first theca the strips are seen to have joined up ventrally with the usual zig-zag suture. Dorsally they are attached to the apical part of the sicula (Text-fig. 2J) and when growth continues beyond this, meet the nema, always extending regularly upwards along it in an oblique direction.

INTERTHECAL SEPTUM OF *MONOGRAPTUS DUBIUS*

It can be seen in Text-fig. 2G that the base of the first interthecal septum is formed by the first individual, since the theca has grown up as a complete tube above the lower part of the second theca. Yet in the distal part it can be seen that the growth-lines are continuous

from it to the second theca, and, being closely spaced, do not correspond with the widely spaced lines on the distal part of the first theca (Text-fig. 2J). Thus this part was formed by the second individual, and there must have been a change-over somewhere along the length of the septum.

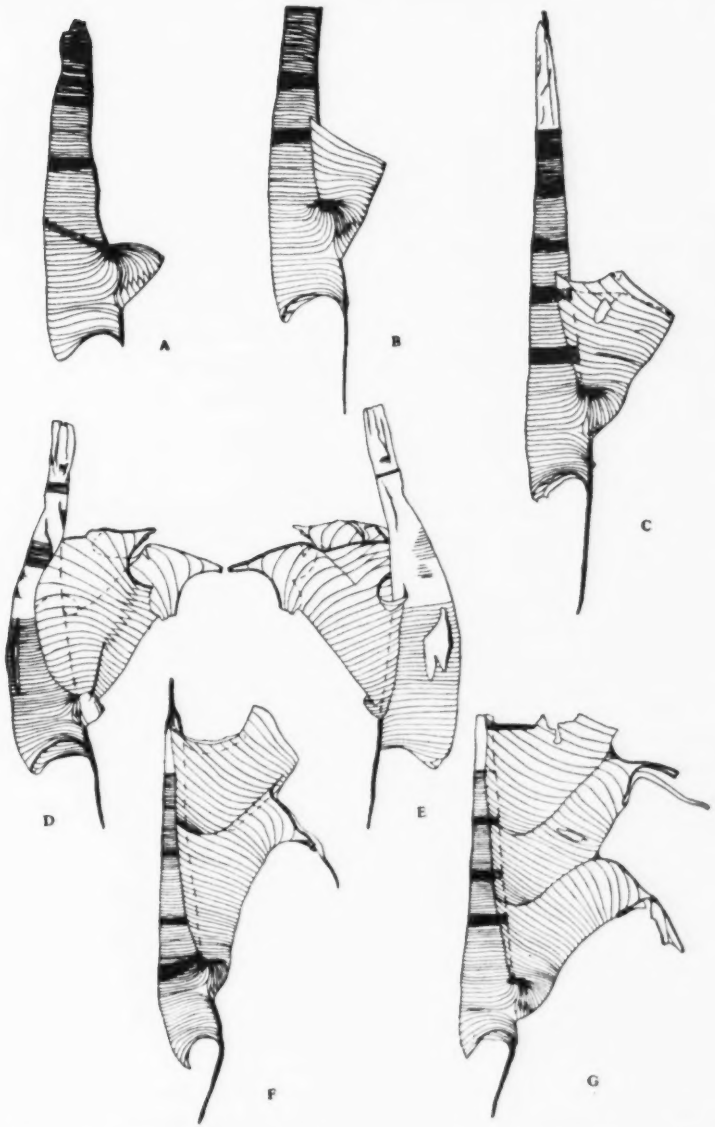
Text-fig. 3A shows a fragment from a later part of the colony; it consists of the ventral wall of a theca with its thickened margin at the top, and its downward continuation as the interthecal septum of the previous theca. This shows the way in which the change occurs, by an interdigitation of chitin strips produced by the two individuals. At the base the strips bending round from the wall of the earlier theca appear to meet in the middle; further up they fail to do so, showing as triangular areas at the sides of the septum. Between these some wedge-shaped strips are intercalated. At first they occur only in the middle region, but then become bigger until they reach the sides; finally they continue on to the outer wall of the rhabdosome, and this occurs before the aperture of the earlier theca is passed (Text-fig. 3B).

That this arrangement occurs in the first interthecal septum as well as later ones can be seen in one individual where only th 1 and the base of th 2 are present, with a part of the interthecal septum broken off and showing a similar pattern of central wedges and triangular pieces at the sides (SM. A24359). In other fragments, however, the arrangement, although essentially the same in plan, forms a much less regular pattern, and it is clear that there can be considerable variation. There is some indication that this interdigitation may be less or even almost absent in one or two specimens, although there is insufficient material to study this variation closely. At any rate the method described above is probably the main way in which the interthecal septum is formed.

Identification of *Monograptus dubius*

The dimensions and general appearance of the form whose development is described above agrees with the description of *M. dubius* given by Lapworth, Elles, and Wood (1901–1918, p. 376). Among the specimens obtained, however, were a few (not figured here, SM. A24360 to A24363), which although the same as these in outward form, showed a somewhat different type of development when cleared preparations were made. The main differences are that in the latter forms there are no "Stillstandsgürtel" on the metasicula, and the initial bud shows no lateral thickened bar and hence no unconformity on the reverse side. Apparently these early stages only are affected, but little is known of the interthecal septa of the two forms and there might also be some variation here.

The type described in full here, with "Stillstandsgürtel" on the



TEXT-FIG. 4.—Growth-stages of *Monograptus chimaera*. A, initial bud; B–E, first theca; F, second theca stage; G, third theca stage. All reverse side except E. A–E × 30 approx.; F, G × 24 approx. SM. A24364 to 24369.

metasicula and a thickened bar on the first theca, is essentially like the specimen described by Kraft (1926, p. 232; fig. 4, p. 237), which was possibly *M. dubius*. The other form, of which there are only a few specimens in this assemblage, is the same as the *M. dubius* described by Wiman (1893) and Cox (1934).

The forms can only be distinguished in cleared preparations when these can be made. Although both are at present referred to *M. dubius*, it should be realized that there must be some uncertainty in identification since only specimens with a few thecae are available.

Development of *Monograptus chimaera*

Initial bud and first theca.

The bud of *M. chimaera* begins, as does that of *M. dubius*, with a few short strips of chitin laid along the edge of the foramen in the sicula. Succeeding growth bands are longer, gradually extending right along the foramen margin and consisting of alternate strips of chitin joining in the middle with a zig-zag suture. The upper strips, however, instead of forming a transverse thickened bar as in *dubius*, pivot round a much smaller thickened region quite close to the virgella, and so grow round and up the reverse face of the sicula without any such pronounced unconformity as is seen in *dubius* (Text-fig. 4A, B). On the obverse side the bands are attached to the wall of the metasicula and the theca grows upwards without unconformity.

The space for the second theca is formed as in *dubius* when the dorsal edges of the chitin strips grow free away from the sicula (Text-fig. 4C). Here about four bands are usually involved before the succeeding strips join dorsally as well as ventrally, forming the base of the first interthecal septum. About five growth-bands are added to the theca in this way.

The characteristic shape of the first theca is produced by the addition of a few strips which are attached to the next theca dorsally but separate at the ventral edge, thus forming a ventral apertural notch (Text-fig. 5B). One or two of the last strips are also free dorsally, becoming successively shorter, and the theca terminates in a pair of short lateral spines (Text-fig. 4D, E, F).

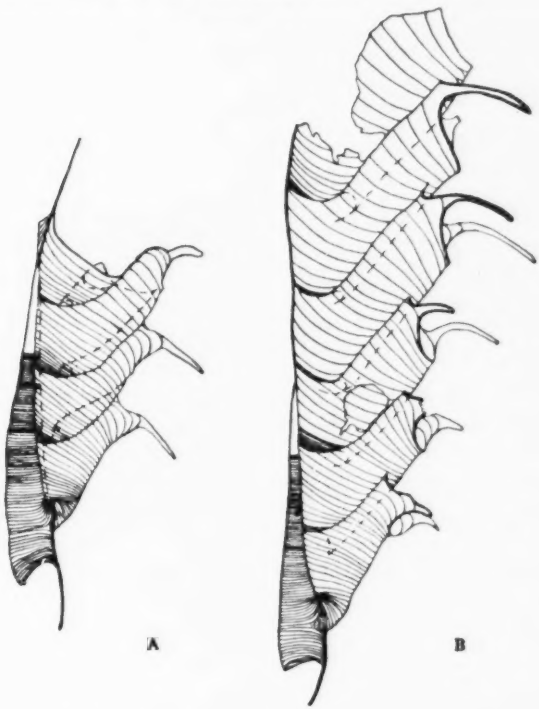
The second theca.

As in *dubius* the early chitin strips of this theca are usually very narrow, and the first few are attached across the thickened ends of the four " free " bands so that there is an unconformity between the two thecae. Later strips extend on each side from the sicula to the dorsal part of the first theca, where it has formed a complete tube; they are wider and join ventrally with a zig-zag suture when the aperture of the first theca is passed.

In exactly the same manner as the first theca, this second one grows at first up the metasicula, then has four to five " free " bands, and finally produces a complete tube comprising the wall of the theca and the interthecal septum (Text-fig. 4F). It is completed by the formation of a ventral notch and apertural spines but here only two or three growth-bands are involved in the formation of the notch, compared with four or five in the first theca, and none of the growth-bands are free at their dorsal edges (Text-fig. 4G).

Successive thecae.

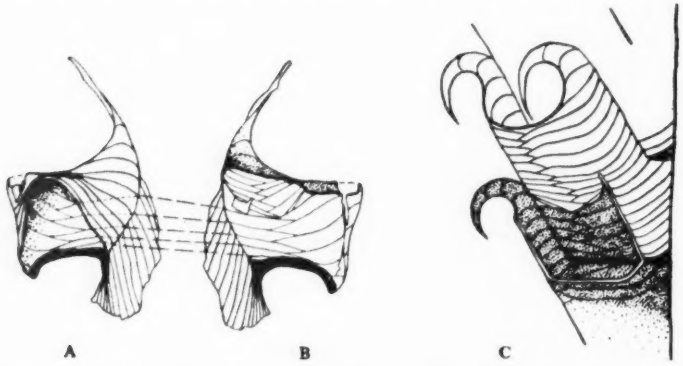
These all grow out from the previous one in a way similar to that described above, and all have an unconformity at the base. Later thecae have more widely spaced growth-rings and they become larger so that the width of the rhabdosome as a whole increases. However, the number of rings per theca (after the first two, which have more) appears to be more or less constant, about 16, and the inclination is



TEXT-FIG. 5.—Later growth-stages of *Monograptus chimaera*. A, four theca stage; B, six theca stage. Both reverse side × 20 approx. SM. A24370 and 24371.

constant at 40°. The ventral notch decreases until in the fourth and later thecae there is little trace of it, but the spines become more prominent and arise on each side from the dorsal part of the edge of the theca, extending laterally and usually bending slightly downwards (Text-fig. 5B).

In these preparations the true *M. chimaera* and its variety *salweyi* cannot be distinguished with certainty. The chief difference is in the size, *salweyi* being the smaller in length and width (Lapworth, Elles, and Wood, 1901–1918, p. 400). If this criterion, particularly of width, is applied, then the majority of the specimens figured are probably



TEXT-FIG. 6.—Interthecal septum of *Monograptus chimaera*. A, B, drawing of one theca from the ventral and dorsal sides showing the growth-lines on the interthecal septum. × 35 approx. SM. A24372. C, diagram illustrating the arrangement seen in A and B in place in the rhabdosome. The ventral and left-side walls of the earlier theca are cut away and the cut edges represented by double lines. The parts belonging to the earlier theca are stippled.

the true *chimaera*; Text-figs. 4B, 4F, and possibly 5A, being somewhat narrower, may be the variety *salweyi*.

INTERTHECAL SEPTUM OF *MONOGRAPTUS CHIMAERA*

In Text-fig. 4D, E, F, an almost complete theca is observed growing above the base of the next which will form, and in Text-fig. 4E the growth-lines on the interthecal septum can be seen; this makes it clear that the base of the septum is formed by the earlier individual. However, an interthecal septum becomes the ventral wall of a theca when the aperture of the previous theca is passed, and in any of the later diagrams (e.g. Text-fig. 5B) the growth-lines on this part are seen to be continuous with those on the lateral wall of the rhabdosome; hence they are produced by the next individual and there must, as in *dubius*, be a change-over along the length of the septum.

The fragment illustrated in Text-fig. 6A, B shows how this change occurs in *chimaera*. About four or five wide growth-bands form the most part of the septum, and are continuous on to the outer wall of the earlier theca. They bear no relation to the narrower bands forming the base of the next theca, of which a fragment is seen. They are rather narrower in the middle, where the two halves overlap, and thus form a concave margin at the top; the two sides of the uppermost band may not meet in the middle, thus adding to this effect. Into this hollow are set a number of narrower bands, making an unconformity with the earlier ones. This unconformity is obviously the place where the next individual takes over formation of the membrane, just before the aperture of the previous theca is passed. The first few strips produced by the later theca must be separate from the strips which form the lateral walls of the rhabdosome, but they become continuous as soon as the hollow mentioned above is filled in (Text-fig. 6C).

Thus in *chimaera* the change-over along the septum occurs at a definite place, with no interdigitation as there is in *dubius*.

Acknowledgments

I wish to thank Dr. O. M. B. Bulman for all the help and advice he has given me in this work, and for passing on to me the material on which it is based, which was originally collected by Dr. W. G. Kühne. I am also indebted to Mr. A. G. Brighton for the loan of some of Cox's specimens from the Sedgwick Museum Collection; to the Council of Girton College for a Tucker-Price Studentship, and to the Department of Scientific and Industrial Research for an award, during the tenure of which the work has been carried out.

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REVIEWS

KORALLEN AUS DEM OBERKARBON (NAMUR) DES OBERSCHLESISCHEN STEINKOHLEN-BECKENS. By O. H. SCHINDEWOLF. *Akad. Wiss. Liter. Mainz (Abhandl. Math.-naturw. Kl.)*. Jahrg. 1952, Nr. 4, pp. 143–227, with 29 figs. and 2 plates.

M. Schwarzbach described as two species of *Zaphrentis* some corals from the Namurian of Upper Silesia in 1937. Professor Schindewolf has re-examined the corals in the fauna of that area and finds that they fall into five genera, of which one is new, and eight species, of which seven are new. He also introduces the new subgeneric name *Pycnocoelia* for *Weissermelia* Schindewolf *non* Lang, Smith and Thomas, but he is misleading in his comment on the type-species (p. 165)—he overlooks his original designation in 1942 of *W. compacta* Schindewolf as the type, so that *W. corpulentum* Schindewolf cannot be ranked as such. There are useful discussions of the Superfamily Zaphrentoidida, and of two of its families, the Zaphrentoididae and the Polycoeliidae. His interpretation of *Fasciculophyllum* Thomson involves the rejection of that of Hill 1940 (for which he adopts *Claviphyllum* Hudson), and the absorption of *Rotiphyllum* Hudson as a synonym. Some of the more generally interesting observations by Schindewolf are concerned with homeomorphy of various stages of the corals. He also shows that the species have a stratigraphical value in the area dealt with. The illustrations are clear and useful.

H. D. T.

METHODS AND PRINCIPLES OF SYSTEMATIC ZOOLOGY. By ERNST MAYR, E. G. LINSLEY, and R. L. USINGER. pp. ix + 328, with 45 figs. McGraw-Hill Book Company, 1953. Price 51*s*.

The recent revival of interest in taxonomy has led to the establishment of one or two University teaching posts in this country and others are under discussion; the publication of this book in the United States suggests that the subject already finds a place in University curricula there. In Palaeontology, as probably in Entomology (and two of the present authors are entomologists), interest in taxonomy has never really lapsed, and this vividly-written textbook on methods and principles is to be welcomed in spite of its high price. The book is primarily designed for students of zoology, but although the problems of the palaeontologist are often different, there is (as the authors note) much common ground of theory and method; and this is a book which should be read by all research students in palaeontology.

The first part of the book is introductory, containing chapters on the species and infra-specific categories, and on classification and the higher categories. Part 2 describes taxonomic procedure, with useful chapters on taxonomic discrimination and taxonomic characters, a brief section on quantitative methods, and a valuable discussion of methods of presentation of results. The "Preparation of Taxonomic Papers" contains helpful suggestions, though the section devoted to illustration is too brief to be of much value and requires a far greater selection of figures. The eight chapters of Part 3 constitute a comprehensive discussion of principles underlying the Rules of Nomenclature. In their preface, the authors state that this section is "open to the criticism that it is an unofficial version of a highly technical and, at the moment, controversial subject": such would be carping criticism, but we might hope that in a second edition of this book a reprint of the Rules of Nomenclature be included. Perhaps too a second edition might contain some reference to Calman's excellent little book, *The Classification of Animals*. References at the end are arranged under chapter headings, and there is a useful glossary.

O. M. B. B.

ZOOGEOGRAPHY OF THE SEA. By SVEN EKMAN. pp. xiv + 417, with 121 text-figs. and 49 tables. Sidgwick and Jackson (Textbooks of Animal Biology), 1953. Price 42*s*.

This work was first published at Leipig in 1935, under the title *Tiergeographie des Meeres*. The English translation has been extensively

revised, and brings together in readily accessible form an immense amount of information on the distribution of present-day marine faunas.

Ekman's classification into regions differs from that of Woodward (familiarized, for example, in Davies's *Tertiary Faunas*) and correlation of the two schemes would have been helped had Ekman provided a small-scale general distribution map. His nomenclature, which may be more scientific, is certainly simpler. The book begins with an account of the warm-water fauna of the shelf seas in the Indo-West-Pacific and Atlanto-East-Pacific ; then follow chapters on the Mediterranean-Atlantic and Sarmatic Regions, the Boreal Fauna of the North Atlantic, the Temperate Fauna of the North Pacific, the Arctic Region, Warm Temperate Faunas of the Southern Hemisphere, and Antiboreal and Antarctic Regions. The last third of the book comprises chapters on Longitudinal Distribution and Bipolarity, the Bottom Fauna of the Deep Sea, and the Pelagic Faunas. There is a bibliography of 600 references.

Palaeontology is by no means ignored, though of course treated purely as a background to present-day faunas; in particular, there is an interesting discussion of the evolution of the warm-water fauna of the shelf seas from the Tethys, and observations on the origin of the Mediterranean, Boreal, and North Pacific faunas. Palaeontological references, however, do not inspire quite the same confidence as the remainder of the book, and do not seem to have been brought so efficiently up-to-date. The work is admirably translated (to judge by its readability) and well produced, and constitutes a valuable addition to the Sidgwick and Jackson *Textbooks of Animal Biology*.

FALKLAND ISLANDS DEPENDENCIES SURVEY : SCIENTIFIC REPORTS 1–5. H.M. Stationery Office, 1953.

The first five of a series of scientific reports of the recent survey of the Falkland Islands Dependencies have just been issued. These comprise Organization and Methods, by V. E. Fuchs (No. 1, price 4*s*. 6*d*.) ; the Upper Cretaceous Cephalopod Fauna of Graham Land, by L. F. Spath (No. 3, 60 pp. with 13 plates, price 30*s*.) ; Lower Cretaceous Gastropods and Lamellibranchs from Alexander I Land, by L. R. Cox (No. 4, with 2 plates, price 5*s*. 6*d*.) ; and Fossil Penguins of Seymour Island, by B. J. Marples (No. 5, with 2 plates, price 5*s*. 6*d*.). The fifth is zoological (No. 2, New Method of Age Determination in Mammals).

LA MER : Ouvrage publié sous la direction de V. ROMANOVSKY, C. FRANCIS-BŒUF, and JACQUES BOURCART, with the collaboration of P. Bohé, J. Y. Cousteau, J. Fuega, M. Guierre, and J. Peytel. pp. 503 (4to), with 16 colour plates and 870 illustrations. Libraire Larousse (Paris), 1953. Price £5 10*s*.

This is a profusely illustrated popular encyclopaedia of the sea, covering all its aspects from early legends and mysteries to tides, waves, and currents, from warships to marine life, from meteorological stations to diving and all forms of under-water exploration, or from yachting to marine deposits. The text is in double column on the pages and the illustrations, averaging two per page through most of the book, are so disposed as to need no numbering or text references. Though popularly written, it is authoritative and seems to be remarkably up-to-date in all sections—the extraction of magnesium from seawater, the Kon-tiki Expedition, and even the disastrous Dutch floods of 1953, all find a place, to mention a few random examples. The illustrations are superb; photographs of marine invertebrates and larval forms by Douglas P. Wilson are particularly striking, though it seems unfair to single out one from the innumerable sources drawn upon. There is unfortunately no index, but a fairly full "Table des Matières". It is a fascinating book one would like to be able to afford as a Christmas present for favoured friends.

O. M. B. B.

LIST OF PUBLICATIONS RECEIVED

Compte Rendu des Seances de la Soc. Serbe de Géol. for 1948 (1950) and for 1949, with parts for 1910, 1913, and 1941 (1953). Belgrade. (Containing numerous short papers on economic mineralogy, Mesozoic and Tertiary stratigraphy, etc.; many with brief English summary.)

Reviews of Research on Arid Zone Hydrology. 212 pp., with 4 folding maps in pocket, 4to, paper bound. Unesco, 1953 (on sale at H.M.S.O.), price 25*s*. (Reports reviewing research in North-West Africa, North-East Africa, Union of South Africa, Middle East, India and Pakistan, United States and Canada, Latin America and Australia.)

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J. E. DUMBLETON. *Wells and Bore-holes for Water Supply, with notes on Water Analysis and Purification*. (2nd ed., revised.) pp. 137, with 39 figs. and frontispiece. London: The Technical Press, Ltd., 1953. Price 17*s*. 6*d*.

R. MCADAM. *Colliery Surveying*. pp. vii + 146, with 80 figures. Oliver and Boyd, 1953. Price 12*s*. 6*d*.

H. D. MACGINITIE. *Fossil Plants of the Florissant Beds, Colorado*. 198 pp., with 75 plates. Carnegie Institution of Washington (Contributions to Paleontology), Publication 599, 1953. Price $5.25 ($5.75 cloth bound).

G. W. TYRRELL. *The Earth and its Mysteries*. pp. xii + 178, with 54 figs. and 14 plates. G. Bell and Sons, 1953. Price 16*s*.

M. R. SAHNI. *Man in Evolution*. pp. x + 272, with 117 figs. Orient Longmans (London: Longmans Green and Co.), 1953. Price 15*s*.

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